Radiometric validation of the Atmospheric Infrared Sounder over the Antarctic Plateau

V. P. Walden, W. L. Roth, R. S. Stone, and B. Halter

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Validation of infrared satellite instruments, after they have been launched into orbit, has traditionally relied on views of the relatively warm sea surface. The Antarctic Plateau provides a complementary validation target from space because it is also homogeneous over large areas, yet it is relatively cold. During summer, cloud-free conditions occur often and the atmospheric humidity is very low with values of total column water vapor around 1 mm. Radiance validation experiments were conducted at Dome C, Antarctica (75°S, 123°E; 3280 m), in January 2003 and again in December 2003/January 2004 for the Atmospheric Infrared Sounder (AIRS). The Polar Atmospheric Emitted Radiance Interferometer (PAERI) measured upwelling and downwelling infrared radiance at the surface during overpasses of NASA’s Earth Observing System Aqua satellite. Radiosondes were launched coincidentally with many of these overpasses. These data are used to calculate the outgoing spectral radiance at the top of the atmosphere (TOA), which is then compared with spectral radiance measurements from AIRS in selected frequency bands. At the 95% confidence level, the brightness temperature differences are $-0.12 \pm 0.24$ K (M-07), $-0.15 \pm 0.23$ K (M-08), and $-0.13 \pm 0.24$ K (M-09). These results validate the AIRS radiances at low temperatures (230 to 250 K) relative to the TOA calculations between 788 and 973 cm$^{-1}$. The cold bias may be partially explained by subvisible or undetected clouds. Ground-based validation experiments over the Antarctic Plateau can yield valuable information about the low-radiance performance of infrared satellite instruments. However, one must use ground-based instruments with a large enough field of view to average over the significant radiance variations that exist across individual snow dunes. Validation of AIRS at low radiance is important because many of the geophysical parameters that are of scientific interest are actually retrieved from low radiance values, including upper tropospheric humidity and properties of high clouds.


1. Introduction

In May of 2002, the National Aeronautics and Space Administration (NASA) launched the Aqua satellite, the second of a series of three Earth Observing System (EOS) satellites. Since that time, numerous field campaigns have been planned for [Aumann et al., 2003; Fetzer et al., 2003] and conducted (see the validation papers in this issue) to provide validation data for Aqua’s various instruments. One of Aqua’s instruments, the Atmospheric Infrared Sounder (AIRS), is an infrared spectrometer, which, because of its fine spectral resolution, is capable of acquiring a wealth of information for weather forecasting and climate studies. To properly validate AIRS radiances, it is necessary to acquire comparable or better data at the surface, both in terms of radiometric uncertainty and spectral resolution.

The sea surface has traditionally been used as a “validation target” for infrared satellite instruments [e.g., Donlon et al., 2002], such as the Moderate Resolution Imaging Spectroradiometer (MODIS). Radiance measurements of the sea surface are used to validate the performance of the heated reference source used to calibrate space-borne radiometers. However, the sea surface is relatively warm (greater than 273 K) despite the fact that many geophysical parameters retrieved from satellite data are measured at low radiance levels. Validation of infrared satellite instruments over the ocean must also contend with the uncertainty induced by large corrections that must be applied to ground-based data to account for humid, marine atmospheres. It is desirable to validate infrared satellite instruments over their full range of operating conditions to ensure proper function and accuracy.
We have selected the Antarctic Plateau as a complementary validation site for infrared satellite instruments and, in particular, for radiometric validation of AIRS. Validation data from the Antarctic Plateau have the potential to reduce the uncertainty in the comparisons of top-of-the-atmosphere (TOA) radiances because the atmospheric correction is small compared with that for more humid and lower-elevation sites. The temperature of the snow surface on the Antarctic Plateau ranges from about 230 K to 250 K in summer at interior stations such as South Pole, Vostok, Dome Fuji, and Dome C, and routinely reaches 200 K to 210 K in winter. Certain areas of the Plateau can serve as a uniform “validation target” in the infrared because they (1) are relatively smooth and level, (2) have high infrared emissivity, and (3) exhibit uniform temperatures over spatial scales typical of the ground-footprint of a satellite instrument (1 to 15 km). In addition, the total column water vapor amount in the atmosphere is so low (around 1 mm in summer) that measurements of upwelling infrared radiation made from the surface (in certain spectral bands), are very nearly equal to those made at the top of the atmosphere. The Plateau (e.g., South Pole Station) also has low values of fractional cloud cover (50–60%) in summer [Town et al., 2005]. An additional advantage is that cross-track scanning instruments on polar-orbiting satellites, such as AIRS and MODIS, view high-latitude sites multiple times each day. Because most polar-orbiting satellites are flown in inclined orbits, it is most advantageous to have a validation site that is not poleward of about 80° so that the satellite will fly directly over the site at certain times. For this reason, South Pole Station is not an ideal validation site. Dome C, however, is located at a latitude that is useful for validating AIRS measurements at radiance levels that are of primary importance to the underlying science of the Aqua mission.

This paper describes AIRS validation experiments that were conducted at Dome C, Antarctica (75°S, 123°E) in the austral summers of 2002/2003 and 2003/2004. The field programs are described, along with the ground-based measurements that were made at Dome C. The methodology for radiometric validation is then presented, including the assumptions that are made in calculating TOA radiances from ground-based measurements. Finally, AIRS radiances measured over Dome C are compared with TOA radiance calculations. A detailed uncertainty analysis is presented that, in general, illustrates the usefulness of the Antarctic Plateau as a validation site for infrared satellite instruments, where radiance comparisons can be made to within a few tenths of a Kelvin.

2. Field Experiments

Two field experiments were conducted at Dome C, Antarctica for AIRS validation, one from 13 to 29 January 2003 and another from 15 December 2003 to 29 January 2004. Dome C is operated for scientific activity jointly by the French Institut Polaire Français Paul Emile Victor (IPEV) and the Italian Programma Nazionale Ricerche in Antartide (PNRA). Dome C is located in East Antarctica, high on the Antarctic Plateau (3280 m above sea level); the Antarctic Plateau is sometimes defined to be the portion of the Antarctic that is 2500 m above sea level. Because of Dome C’s latitude, Aqua passes over Dome C six times each day (occasionally seven), including three overpasses during ascending orbits and three during descending orbits about 8 hours later. The area used for validation in this study (roughly 50 km by 50 km centered over Dome C) is extremely flat with maximum elevation changes of only 20 m [Capra et al., 1994]. There is a significant diurnal cycle in the surface temperature in summer (under clear skies) at Dome C of about 20 K (from 230 K to 250 K), driven by changes in the solar zenith angle; during our field experiment, the solar zenith angle varied from about 53° at noon to 84° at midnight.

Our validation field site was located 1 km west of the summer station at Dome C, and about 800 m from the new Concordia Station that was under construction at the time of this project. Figure 1 shows a picture of the field site. This location was chosen in the “clean area” of Dome C and was never downwind of the station, thus avoiding any pollution from the station’s power generator. Traffic by motor vehicles was minimized at the field site. The area viewed by our ground-based radiation instruments was of pristine snow, which had received no foot or vehicle traffic. During both field seasons, the snow surface consisted of wind-blown snow drifts, called sastrugi. Four surveys of undisturbed snow performed in January 2004 showed that the sastrugi had horizontal dimensions of about 1 to 2 m long and about 30–50 cm wide with vertical dimensions (heights) of less than ±8 cm; the standard deviation of the vertical measurements is 2.5 cm (D. Six, personal communication, 2004). Thus a typical aspect ratio (largest horizontal length:vertical length) is probably about 10:1 to 25:1. Note that these measurements are representative of Dome C during January 2004, but will change in time and also with location across the Antarctic Plateau.

Two towers were erected at the field site in December 2002 and January 2003. A 6-m tower was used for validation in January 2003. A 32-m tower was also constructed during this field season for a separate experiment being conducted by the University of Washington. The Polar Atmospheric Emitted Radiance Interferometer (PAERI) was mounted 24 m above the surface on this tower in December 2003 and January 2004.

3. Data

Data from various instruments were used to characterize the scene observed by AIRS, including infrared measurements from the PAERI and radiometers, as well as from MODIS satellite imagery. Radiosondes were used to characterize the atmosphere.

3.1. Ground-Based Data

Various instruments were deployed during both field experiments. The PAERI was used to measure the upwelling and downwelling spectral infrared radiance. The PAERI is one of several AERIs that were built at the Space Science and Engineering Center (SSEC) at the University of Wisconsin-Madison. Therefore the PAERI’s calibration is quite similar to those AERIs that are currently deployed at the Department of Energy’s Atmospheric Radiation Measurement (ARM) sites [Knuteson et al., 2004a, 2004b] and by the University of Miami [Minnett et al., 2001]. The PAERI has a front-end optical assembly similar to the marine AERIs that
allows the instrument to view at any angle from nadir to zenith, thus viewing the surface and the atmosphere. A standard AERI detector was used at Dome C with a spectral range of 475 to 3000 cm\(^{-1}\) (3 to 21 micrometers) in two bands; channel 1 from 475 to 1800 cm\(^{-1}\) (5.5 to 21 \(\mu\)m) and channel 2 from 1800 to 3000 cm\(^{-1}\) (3 to 5.5 \(\mu\)m). Because the sun was continually above the horizon for the entire duration of both validation experiments, only the channel 1 data are used in this study to avoid the contribution from solar infrared detected in channel 2. Extended range detectors, which are sensitive to around 400 cm\(^{-1}\), are often used in the polar regions because they detect radiance in the semitransparent region of water vapor from 400 to 550 cm\(^{-1}\), but at the expense of detector sensitivity. Because the bandwidth of the AIRS instrument is only from 650 to 2650 cm\(^{-1}\), the extended range detector was not necessary. Therefore a standard detector was used at Dome C to maximize the signal-to-noise of the PAERI.

[11] The AERI design is described by Knuteson et al. [2004a, 2004b]. The interferometric calibration of Revercomb et al. [1988] is used. Best et al. [1997, 2003] describe the infrared sources used by the AERI systems and state that the absolute uncertainties (3\(\sigma\)) for the source temperature and cavity emissivity are 0.10 K and 0.001, respectively. The PAERI uses two infrared sources, one heated and one at ambient temperature, to calibrate a third scene view. The PAERI performance has been characterized regularly through various tests performed at the SSEC. Before the first field season, the blackbody thermistors and electronics rack were both calibrated. In addition, tests to characterize the front-end alignment, radiometric nonlinearity, and the overall radiance calibration were performed in the SSEC laboratory.

[12] In addition, two tests were performed at the beginning of each field season at Dome C to verify the calibration, one inside our field hut and one outside. These tests included viewing a third infrared source that was controlled at an intermediate temperature between the heated and ambient blackbodies. The calibration tests were performed while viewing the third infrared source at 318 K inside the hut and then at 253 K outside. These measurements were then compared to theoretical values derived from knowledge of the source’s temperature and spectral emissivity. The PAERI calibration was verified to be within 0.03 K at 318 K. The measurements at low radiance showed a slight offset compared to the lab measurements, but were still within about 0.04–0.05 K at 253 K. The low-radiance calibration is especially significant because the radiance

Figure 1. Image of the Dome C field site used for AIRS validation. This site is located 1 km west of the summer station. The PAERI is visible on the tower at about 24 m above the snow surface. The PAERI viewed undisturbed snow on the far side of the flags on the surface to the right of the tower.

Figure 2. Infrared brightness temperature spectra of (a) upwelling emission from the surface and (b) downwelling emission from the atmosphere measured by the Polar Atmospheric Emitted Radiance Interferometer (PAERI) at Dome C, Antarctica, on 23 December 2003. The vertical scales on the two plots are different, indicating that the upwelling emission from surface has much less spectral variation than the downwelling emission from the atmosphere.
conditions of the source in this particular test are similar to the actual radiances being measured for AIRS validation, that is, a snow surface with high emissivity at low temperatures. Therefore the absolute calibration of the PAERI measurements in the field is excellent and is comparable to other AERI instruments [Knuteson et al., 2004a, 2004b; Minnett et al., 2001].

[13] Figure 2 shows sample brightness-temperature spectra measured by the PAERI of upwelling radiance from the surface (Figure 2a) and downwelling radiance from the atmosphere (Figure 2b). The brightness temperature spectrum of upwelling radiation shows little variation (<1 K). The brightness temperature differences between 800 and 1200 cm\(^{-1}\) are due primarily to spectral variations in the emissivity of the snow.

[14] The brightness temperature spectrum of the downwelling radiance was taken during clear-sky conditions and shows the spectral signatures of various atmospheric gases such as carbon dioxide around 667 cm\(^{-1}\), ozone between 950 and 1050 cm\(^{-1}\), and methane, nitrous oxide and water vapor at wave numbers greater than 1200 cm\(^{-1}\). Noise was filtered from the PAERI radiances using the principal component method described by Antonelli et al. [2004].

[15] Two infrared thermometers (IRT) (Heimann KT-19 narrowband radiometers; 700–1250 cm\(^{-1}\) (8–14 μm)) were used during both field experiments to help understand how the point measurements made by the PAERI relate to the larger field of view of the AIRS instrument. Data from MODIS were also used for this, as explained below. One sled-mounted platform, NOAA’s Earth System Research Laboratory AIRS Mobile Observing System (referred to as AMOS), was usually pulled behind a snowmobile to make radiometric scans of surface temperature using an IRT. The development and use of this system are described by Maslanik et al. [1999] and Schnell [2004]. A second IRT was operated by U. Idaho (in collaboration with Stephen Hudson from U. Washington) from 30 m above the ground on the tall tower. The tower instrument was able to view about 225° in azimuth from NNE to SSE, and at five different zenith angles (15°, 30°, 45°, 60°, and 72°).

[16] Vaisala radiosondes were launched routinely from Dome C to characterize the atmosphere. Vaisala RS80-GH sondes were used during the first field season and RS90-AG sondes during the second. The radiosondes were launched coincidentally with the two near-nadir overpasses made by AIRS each day. (The near-nadir overpasses were always bracketed by two off-nadir overpasses (with AIRS viewing angles of 40° to 50°), which occurred about 100 min before and after the near-nadir overpass.) To minimize problems with the response time of the radiosonde’s pressure, temperature, and humidity sensors [Hudson et al., 2004], the sondes were stored and prepared at ambient outside temperature [Walden et al., 2005]. The uncertainties in the temperature and water vapor given by Vaisala are ±0.5 K and ±5%, respectively. Figure 3 shows representative temperature and humidity profiles measured by a sonde. The total column water vapor amount for this day was about 1.3 mm. (Even though the atmosphere is extremely dry, AIRS has sufficient sensitivity (10 ppmv) to measure fluctuations of this type in water vapor.) The radiosonde data were supplemented with vertical profiles of ozone, nitrous oxide, methane, CFC-11, and CFC-12, derived from the 15 January 1992 case used by Walden et al. [1998]; these profiles were included in the radiative transfer calculations (described below) for completeness, but do not contribute significantly to the portion of the spectrum where the TOA radiances are compared.

3.2. Ancillary Data From MODIS

[17] Figure 4 displays maps of brightness temperatures at 11 μm (band 31) from MODIS [e.g., Barnes et al., 1998]. Figure 4a shows the TOA radiance over the Antarctic continent, while Figure 4b shows the local area around Dome C. In general, the MODIS brightness temperatures indicate differences in topography over the Antarctic Plateau. Relatively higher temperatures (263 to 268 K) are seen...
around the periphery of the continental ice sheet, while lower temperatures (about 240 K) are observed over the high plateau. Some of the lowest temperatures (below 240 K°C) indicate emission from the tops of clouds.

Figure 4b shows that Dome C is located in a region with fairly homogenous temperatures. It also shows the approximate locations of nine AIRS fields of view that are used for radiance validation. MODIS data are used in this study to provide an estimate of the uncertainty in using the point measurements made by the ground-based radiation instruments to represent the field of view of the nine AIRS measurements. The MODIS MYD021KM data (MODIS/Aqua Calibrated Radiances 5-Min L1B Swath 1km) are used here.

3.3. AIRS Data

Before arriving at Dome C for each of the two field seasons, predictions of the timing of the AIRS overpasses were downloaded from NASA’s Earth Observatory Satellite Overpass Predictor (http://earthobservatory.nasa.gov/MissionControl/overpass.html). This allowed our field crew to synchronize both the PAERI measurements and radiosonde launches to the overpasses of the Aqua satellite. The acquisition of PAERI data was typically started before the first overpass in a sequence of three ascending or descending orbits. However, the PAERI viewing angle was always set to match the second, near-nadir overpass in a sequence.

The AIRS level 1B radiances (version 3.2.7) are used to compare with the TOA radiance calculations described below. (These radiances were originally obtained from the AIRS Validation Team data archive at NASA’s Jet Propulsion Laboratory and are now available from the Goddard Earth Science Data and Information Services Center (GES DISC) at http://disc.gsfc.nasa.gov/AIRS/airsL1B_Rad.shtml.) The spectral radiances from nine contiguous AIRS fields of view (FOV) (in a $3 \times 3$ array; Figure 4b) are averaged and then used for each spectral radiance comparison, with Dome C located within the center FOV. This was done to reduce the noise in the AIRS measurements. The heat generated by the station itself has a negligible effect on the overall radiance detected by AIRS over this region; elimination of the center FOV changes the average radiance over this area by less than 0.05 K.

Version 6.7.0 (19 November 2003) of the AIRS channel property file was used for this study. The radiance comparisons discussed below use a stringent selection of particular channels (wave numbers) that have excellent noise performance. A subset of 1738 channels was used here for radiance validation (out of a total of 2378) that were suggested by the Atmospheric Spectroscopy Laboratory at the University of Maryland-Baltimore County (UMBC) (L. Strow and S. Hannon, personal communication, 2005). This subset was more restrictive than simply using the CalChanSummary and ExcludedChans flags in the AIRS metadata as suggested by Olsen et al. [2005]. Subsetting these channels further by “AB state” of the AIRS detectors has a negligible effect on the overall results of this study.

4. Methodology for Radiometric Validation

The upwelling radiance reaching an instrument at any level within the atmosphere (including TOA) is given by:

$$L_{u,i}(\theta, \phi) = \tau_{u,i}(\theta, \phi) \left[ e_{u,i}(\theta, \phi) B_e(T_e) + \int_0^{2\pi} \int_0^{\pi/2} f_{u,i}(\theta, \phi, \theta_i, \phi_i) L_{u,i}(\theta, \phi) \cos \theta_i \sin \theta_i \, d\theta_i \, d\phi_i \right] + E_u(\theta, \phi)$$

Figure 4. Maps of MODIS brightness temperatures (band 31) at 11 μm measured on 23 December 2003 near 0845 UTC (a) over the Antarctic continent and (b) over Dome C. In both panels, the star denotes the location of Dome C Station. Figure 4b shows the approximate locations of nine AIRS fields of view that are used for radiance validation. The area shown is approximately 50 km by 50 km. The brightness temperatures are given in Kelvin; the color scales for the two figures are different. The range of brightness temperatures in the AIRS fields of view shown in Figure 4b is about 0.7 K.
Figure 5. Conceptualization of our approach to radiance validation over the Antarctic Plateau. The numbers represent the different terms in the radiative transfer equation in the text (equation (1)). Term 1 is emission from the surface, term 2 is emission from the atmosphere that is reflected off the surface and into the instruments, and term 3 is emission from the atmosphere. The Polar Atmospheric Emitted Radiance Interferometer (PAERI) measures the emission from the surface. Radiosondes are used to characterize atmospheric properties needed for radiative transfer calculations (e.g., vertical profiles of temperature and humidity).

where $L_{\text{u},1}$ is the upwelling spectral radiance at a given pressure level, $\tau$ is the atmospheric transmission between the surface and the instrument, $\varepsilon$ is the spectral emissivity of the surface, $B_s(T_s)$ is the Planck radiance at the surface temperature ($T_s$), $f_i(\theta_i, \phi_i, \theta, \phi)$ is the bidirectional reflectance distribution function (BRDF) of the snow, $L_{\text{d},1}$ is the downwelling spectral radiance from the atmosphere, $\theta_i$ and $\phi_i$ are the zenith and azimuthal angles of incidence of atmospheric radiation to the surface, $E$ is the emission from the atmosphere between the surface and the instrument, and $\theta$ and $\phi$ are the zenith and azimuthal viewing angles of the instrument. The first term in equation (1) is the emission from the surface to the instrument. The second term is the downwelling emission from the atmosphere that is reflected off the surface and into the instrument. The third term is the emission from the atmosphere itself between the surface and the instrument. The first two terms are multiplied by the atmospheric transmission between the surface and the instrument.

[23] Figure 5 shows a conceptualization of our validation approach. This approach derives accurate values of the surface emission from the PAERI data (i.e., $\varepsilon_{\text{s}},$ and $T_s$), then uses a radiative transfer model built especially for the AIRS project (k-Compressed Atmospheric Radiative Transfer Algorithm; kCARTA [Strow et al., 1998]) to calculate the TOA radiance. This method works quite well for the ground-based measurements obtained at Dome C. Reflection of downwelling atmospheric emission off the snow surface is quite small in the atmospheric window region (800 to 1200 cm$^{-1}$, excluding 1000 cm$^{-1}$ ozone emission). This is because the snow is very nonreflective in this spectral region since the emissivity is nearly equal to unity. Also, the downwelling emission (under clear skies) is nearly zero, and the transmission of the atmosphere is nearly unity. For the ground-based PAERI measurements between 800 and 1000 cm$^{-1}$, the second and third terms of equation (1) can be neglected, and $\tau_v$ can be assumed to be unity in the window region. Therefore the PAERI measures the emission from the surface ($[\varepsilon(V, \phi)B_s(T_s)]$ directly. 

[24] To calculate the TOA radiance using kCARTA, it is necessary to specify the spectral emissivity and surface skin temperature separately, because kCARTA requires both of these parameters as input. Note that a reasonable approximation (to say $\pm 0.02$) of the spectral emissivity is sufficient for the conditions measured at Dome C, as long as the product $\varepsilon_{\text{s}}B_s(T_s)$ is consistent with the PAERI measurements. In other words, the radiative transfer model uses $\varepsilon_{\text{s}}$ and $B_s(T_s)$ separately to calculate $\varepsilon_{\text{s}}B_s(T_s),$ but for accurate validation, this calculated product must be equal to that measured by the PAERI.

[25] The subsections below describe the steps taken to provide radiance validation for the AIRS instrument.

4.1. Overpass Selection Criteria

[26] Three criteria were used to select the optimal conditions under which to perform radiance validation. First, it was essential to use PAERI data from the 24-m tower rather than the 6-m tower to maximize the PAERI footprint on the snow surface. Therefore only data from the second field season were used. Data were used only when the PAERI was in rapid-scanning mode to provide adequate spatial coverage of the snow surface in a relatively short amount of time (15 min).

[27] Second, clear-sky conditions were determined from visual observations, downwelling radiances measured by the PAERI, and by examining histograms of MODIS data in the vicinity of Dome C. Figure 6 displays a time series of downwelling radiances at 900 cm$^{-1}$ for the entire second field campaign. Clear-sky conditions exist when the downwelling radiance is nearly equal to zero. The mean downwelling radiance between 800 and 850 cm$^{-1}$ (a transparent microwindow that is sensitive to cloud cover) is less than 1 mW m$^{-2}$ sr$^{-1}$ (cm$^{-1}$)$^{-1}$ for each overpass used in this study. To ensure that clouds were not present over the larger area of the AIRS measurements, histograms of MODIS data were generated for all pixels within approximately 50 km of Dome C. These histograms showed a very small range of temperatures (minimum of 0.4 K, maximum of 1.6 K) with no outliers.

[28] Third, AIRS data were used only if a radiosonde was launched within two hours of the corresponding overpass. This particular time limit was chosen to minimize changes in atmospheric emission, but also to include the two off-nadir overpasses that bracketed the near-nadir overpass when the sonde was launched. Since sequences of three overpasses occurred twice each day, we typically launched two sondes per day.
entire atmosphere is equal to that measured by the PAERI at zenith and that the snow is a Lambertian surface. Both of these are normally poor assumptions, but it is acceptable to assume them here because we are most interested in the product of $e_u(\theta)B_u(T_u)$, not the emissivity specifically. Using these assumptions, the spectral emissivity can be determined from

$$e_u(\theta) = L_{s,1}(\theta) - L_{s,1}(0^\circ)$$

(2)

where $L_{s,1}(\theta)$ is the upwelling radiance and $L_{s,1}(0^\circ)$ is the downwelling radiance at zenith, both measured by the PAERI. As described by Bower [2001], the spectral emissivity and surface temperature are derived simultaneously by adjusting the surface temperature $T_s$ until the derived spectral emissivity varies smoothly as a function of wave number. We used the spectral range from 757–782 cm$^{-1}$ to determine the smoothly varying emissivity.

[32] Most of the viewing angles of the PAERI measurements are not exactly matched to the AIRS viewing angle. Because of this, the emissivities derived from the PAERI are adjusted to match the emissivity that would be measured at the AIRS viewing angle. This is done by calculating a ratio of emissivities using the theoretical model for snow emissivity from Dozier and Warren [1982]. The adjusted emissivity at the AIRS viewing angle is then calculated by multiplying the original emissivity derived from the PAERI measurements times the ratio of emissivities calculated from the Dozier and Warren model.

$$e_u(\theta)_{AIRS} = e_u(\theta)_{PAERI} \times \frac{e_u(\theta)_{AIRS}}{e_u(\theta)_{PAERI}}$$

(3)

In other words, the Dozier and Warren model is used here to simply scale the emissivity derived from the PAERI to the desired AIRS viewing angle.

### Table 1. List of Overpass Times Used for Radiance Validation Comparisons

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<thead>
<tr>
<th>Overpass Time, UTC</th>
<th>AIRS Granule Number</th>
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<td>079</td>
</tr>
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</table>

*The overpass times in regular type occurred in the afternoon, while those in italics occurred in the late evening; local time = UTC + 8 hours. The AIRS granule numbers are also shown. Times are in yyyy.mm.dd:hr:mn format, where yyyy is year, mm is month, dd is day, hr is hour, and mn is minute; read 2003.12.23:08:46 as 23 December 2003, 0846 UTC.*
4.3. TOA Radiance Calculation

The values for $e_u(\theta_{\text{AIRS}})$ and $T_s$ are then used as input into the kCARTA radiative transfer model to calculate the TOA radiance over Dome C at the AIRS viewing angle. The radiosonde data are used to construct a model atmosphere of temperature and water vapor versus pressure for each of the sixteen overpasses. Each of these model atmospheres is combined with the twelve corresponding values of $e_u(\theta_{\text{AIRS}})$ and $T_s$ that were derived from the individual PAERI measurements to create 192 cases for input into kCARTA.

Because the atmosphere is so cold and dry over Dome C, it contributes very little to the TOA radiance. Uncertainties in the radiosonde data induce a negligible error in the calculated TOA radiance. Figure 7 shows the results of a sensitivity study of the calculated TOA radiance (Figure 7a) and the difference in the TOA brightness temperature (Figure 7b) with and without uncertainties of $+0.5$ K in temperature and $+10\%$ in relative humidity with respect to water (RHw). (Note that the uncertainty in relative humidity used here is twice the value stated by Vaisala (5\%) to account for the slow response of the Vaisala humidity sensor at low temperature.) Figure 7b shows that the error induced by radiosonde errors is $0.01$ K or less throughout much of the window region (800 to 1150 cm$^{-1}$). The frequencies where the error is small ($<0.01$ K) are the most ideal for radiance validation.

Figure 7 also shows the band limits of various AIRS detector modules, which are chosen here for radiance validation. These bands are M-07 ($910.3–973.2$ cm$^{-1}$; $10.275–10.985$ μm), M-08 ($851.0–903.3$ cm$^{-1}$; $11.070–11.751$ μm), and M-09 ($788.3–851.0$ cm$^{-1}$; $11.743–12.685$ μm). Note that the M-5 ($1055–1135$ cm$^{-1}$; $8.807–9.479$ μm) and M-6 ($973.2–1045.5$ cm$^{-1}$; $9.565–10.275$ μm) bands are not validated here because we have inadequate information on the vertical profile of ozone.

4.4. Spatial Representativeness of the PAERI Measurements

For Dome C to serve as an adequate site for radiance validation, the accurate point measurements made by the PAERI must represent the larger field of view of the AIRS instrument. During the first field season at Dome C, both the PAERI and AMOS’s IRT measured temperature fluctuations of about 2 K when scanning across a snow dune (sastrugi); the horizontal dimension of this feature was on the order of 1 m. Even though sastrugi in the summertime at Dome C are less than about 15 cm in amplitude, significant temperature gradients were measured across these features because of differing amounts of solar absorption, caused effectively by differing solar zenith angles across a sastrugi. The sunlit side of sastrugi is warmer and brighter than the shaded side, as was documented by both the AMOS’s IRT sensor and its Licor solar radiation sensor (which measured upwelling solar flux).

This was a significant issue during the 2002/2003 field season because the ground footprint of the PAERI was fairly small compared to the horizontal dimension of the sastrugi. The PAERI’s angular field of view FOV is 2.6°, full-width FOV at half maximum [Knuteson et al., 2004a]. Thus the ground footprint of the PAERI in January 2003 was about 0.3 m in diameter when the instrument was...
Table 2. Uncertainty Estimates (in Kelvin) for Scene Variability Measured by MODIS (Column 2), the AIRS Instrument Noise (Column 3), and the Scene Variability Over the Twelve PAERI FOVs (Column 4).

<table>
<thead>
<tr>
<th>Overpass Time, UTC</th>
<th>MODIS</th>
<th>AIRS</th>
<th>PAERI</th>
</tr>
</thead>
<tbody>
<tr>
<td>2003.12.23:08:46</td>
<td>0.17</td>
<td>0.11</td>
<td>0.34</td>
</tr>
<tr>
<td>2003.12.23:15:15</td>
<td>0.37</td>
<td>0.11</td>
<td>0.56</td>
</tr>
<tr>
<td>2003.12.26:15:46</td>
<td>0.29</td>
<td>0.16</td>
<td>0.49</td>
</tr>
<tr>
<td>2003.12.27:06:43</td>
<td>0.16</td>
<td>0.23</td>
<td>0.20</td>
</tr>
<tr>
<td>2003.12.27:08:21</td>
<td>0.10</td>
<td>0.17</td>
<td>0.27</td>
</tr>
<tr>
<td>2003.12.27:13:13</td>
<td>0.18</td>
<td>0.18</td>
<td>0.58</td>
</tr>
<tr>
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<td>0.14</td>
<td>0.34</td>
</tr>
<tr>
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<td>0.23</td>
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</tr>
<tr>
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<td>0.17</td>
<td>0.34</td>
</tr>
<tr>
<td>2004.01.12:08:21</td>
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<td>0.10</td>
<td>0.20</td>
</tr>
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<td>0.12</td>
<td>0.54</td>
</tr>
</tbody>
</table>

The scene variability measured by MODIS is calculated as the standard deviation of band 31 pixels in the nine AIRS FOVs used for validation. The M-8 detector module is shown as representative of errors in the AIRS instrument, because it is typically the channel (of those considered in this study) with the largest measurement uncertainty; the M-07 and M-09 channels are within 0.08 K of the M-08 values. The scene variability measured by the PAERI is calculated as the standard deviation of the measurements over the twelve PAERI FOVs. The overpass times in regular type occurred in the afternoon, while those in italics occurred in the late evening.

5. Validation Results

5.1. Brightness Temperature Comparisons

Figure 8 shows two comparisons of AIRS spectra with kCARTA calculations of the TOA radiance. The agreement is quite good (to about 0.1 K) across the window region from 750 to 1000 cm⁻¹. The large difference in brightness temperature between the two comparisons is due to the diurnal cycle at Dome C; the 0846 UTC case occurred at 1646 local time (LT), while the 1515 UTC case occurred at 2315 LT.

To provide an overall comparison of the AIRS measurements with the TOA calculations, the brightness temperature spectra for AIRS and the TOA calculations are averaged over the bandwidths of the AIRS detector modules (M-07, M-08, M-09). The difference of these averages (AIRS observation minus TOA calculation) is then taken. Histograms of these differences are shown as the solid bars in Figure 9 for the three detector bands. The histograms are normalized and represent the 192 comparisons from method 1. The sixteen comparisons of method 2 are shown as the shaded bars. The mean values for the two histograms in each panel are equal and are listed in Figure 9. The standard deviations for both histograms are given below the mean values in each panel.
tobin et al. made direct radiance
comparisons between AIRS observations and measurements made by the Scanning High-resolution Interferometer Sounder (S-HIS) aboard a high-altitude aircraft over the Gulf of Mexico.

5.2. Uncertainty Analysis

The standard deviations of the brightness temperature differences in Figure 9 are due to a combination of measurement uncertainties in the AIRS and PAERI instruments, as well as scene variability and errors in the radiative transfer calculations. The uncertainty is dominated by both the variability in emission across the surface snow on the spatial scale of tens of kilometers (over the nine AIRS ground footprints) and the small-scale (meters) variability across individual sastrugi. The various contributions to this uncertainty are discussed, in turn, below.

The uncertainty in the AIRS measurements is estimated by simply taking the standard deviation of the nine brightness temperature spectra for each footprint. These uncertainties are listed in column 3 of Table 2 for each measurement.

The uncertainty in the TOA calculations come from several sources, but is dominated by the radiance variability across the relatively small PAERI ground footprint. The variability in the PAERI measurements is estimated as the standard deviation of the brightness temperatures of the twelve individual measurements used for validation during a single AIRS overpass. These uncertainties are listed in column 4 of Table 2 and ranged from about ±0.2–0.5 K for the afternoon overpasses to about ±0.5–1.0 K for the late evening/early morning overpasses when the sun is low in the sky and large thermal gradients develop across the sastrugi. The uncertainty in the PAERI measurements themselves is small (±0.05 K at low radiance values), as explained in section 3.1. The uncertainty in the view-angle emissivity adjustment is estimated to be about ±0.1 K. This assumes that the error in the emissivity calculated from the Dozier and Warren model is ±0.001. The uncertainty in the ratio of emissivities then yields a radiance error of about 0.1 K at temperatures between 230 and 250 K. This is a conservative estimate since one might expect any biases in the modeled emissivities to cancel somewhat when the emissivity ratio is calculated. The uncertainties in the radiative transfer calculation, or more appropriately the atmospheric correction, are small (<0.02 K) as can be seen in Figure 7b. Finally, the standard deviation of the MODIS data across the nine AIRS FOVs is used as the uncertainty in how representative the PAERI measurements are of the larger AIRS FOV; these values are listed in Table 2.

Combining all of the uncertainties in the TOA calculations with the uncertainty in the AIRS measurements gives an estimate of the uncertainty of method 1. Depending on the particular case, the uncertainty in the brightness temperature differences ranges from ±0.3 to 1.1 K, which brackets the standard deviation (0.66 to 0.67 K) of the histograms in Figure 9. This uncertainty is dominated by the variability over the PAERI FOVs. The standard deviations for method 2 (0.42 to 0.45 K) are correspondingly smaller because averaging over the PAERI FOVs reduces the variability of the measurements due to differential heating of the sastrugi.

The difference in the mean values of the AIRS observations and the TOA calculations are estimated using the sixteen cases from method 2. To do this, we calculate the brightness temperature difference (AIRS minus TOA) for each case and then determine the 95% confidence interval of those differences using

\[ \sigma_{95\%} = \pm t_{0.025} \frac{\sigma_2}{\sqrt{N}} \]  

where \( t_{0.025} \) is the student’s critical point for 15 degrees of freedom (N-1), \( \sigma_2 \) is the standard deviation of the temperature differences from method 2, and N equals 16. (This assumes that the differences are normally distributed.) The null hypothesis is that the difference in the mean values of both the AIRS measurements and the TOA calculations is zero (or that the mean values are equal). The values of \( \sigma_{95\%} \) for each detector band are ±0.24 (M-07), ±0.23 (M-08), and ±0.24 (M-09). Thus at the 95% confidence level, the mean values of the brightness temperature differences are –0.12 ± 0.24 K (M-07), –0.15 ± 0.23 K (M-08), and –0.13 ± 0.24 K (M-09). Each of these intervals contain 0 K, which essentially validates the AIRS radiances relative to the TOA calculations at Dome C in each of these detector bands.

5.3. Potential Biases

The primary sources of potential bias in the brightness temperature comparisons are subvisible clouds and

\[ \text{Figure 8. Comparison of two AIRS spectra with top-of-the-atmosphere radiance calculations. Both comparisons are from 23 December 2003, one in the afternoon (0846 UTC) and the other in late evening (1515 UTC). The comparisons are shown in a portion of the “atmospheric window” from 750 to 1000 cm}^{-1} \].
undetected clouds. Although every precaution was taken to screen the cases for clouds, the possibility remains that either subvisible clouds were present over the validation area, or that clouds went undetected in our analysis of the MODIS histograms. Thin layers of ice crystals (sometimes called ice fog or diamond dust) and water fog were observed near the instrument tower during both field seasons at Dome C and are examples of clouds that could potentially bias the comparisons.

The presence of subvisible clouds between the PAERI and AIRS instruments could potentially be responsible for some of the cold bias shown in Figure 9. The effect of subvisible clouds is investigated here through the use of a line-by-line radiative transfer model (LBLRTM) [Clough et al., 1992; Clough and Iacono, 1995] in conjunction with the scattering code DISORT [Stamnes et al., 1988]; we used LBLDIS for this study, which is described by Turner [2005]. First, the clear-sky downwelling infrared spectral radiance was calculated with LBLRTM for each overpass using the radiosonde data as input. A viewing angle of 50° from nadir was used to calculate the maximum effect of a subvisible cloud on AIRS validation. The calculated window radiances were then compared with the PAERI measurements in the transparent microwindow between 829 and 839 cm\(^{-1}\). This spectral band is particularly sensitive to any emission from clouds because there is so little gaseous emission from the atmosphere; the clear-sky radiance is about 0.2–0.4 mW m\(^{-2}\) sr\(^{-1}\) (cm\(^{-1}\))\(^{-1}\). The calculated window radiances compare remarkably well with the PAERI measurements; the differences (calculations – measurements) in the spectral-average radiance over the 835-cm\(^{-1}\) microwindow are less than 0.1 mW m\(^{-2}\) sr\(^{-1}\) (cm\(^{-1}\))\(^{-1}\).

LBLDIS was then used to approximate the optical thickness of a subvisible cloud that would add an additional 0.1 mW m\(^{-2}\) sr\(^{-1}\) (cm\(^{-1}\))\(^{-1}\) to the 835-cm\(^{-1}\) clear-sky window radiance. This was done for two overpasses, one in the afternoon (23 December 2003 at 0846 UTC (1646 LT)) and one in the late evening (23 December 2003 at 1515 UTC (2315 LT)). The temperature profiles for both cases are similar except that the surface temperature is 231 K at 1515 UTC and 244 K at 0846 UTC; a strong near-surface temperature inversion existed at 1515 UTC, but not at 0846 UTC. An ice cloud was positioned at 500 m above the surface in the model, which is near the maximum tropospheric temperature for both temperature profiles. The optical depth of the cloud was then adjusted until the radiance at 835 cm\(^{-1}\) was 0.1 mW m\(^{-2}\) sr\(^{-1}\) (cm\(^{-1}\))\(^{-1}\) above the clear-sky calculation. The infrared optical depth of the resultant ice cloud was 0.003.

The radiative effect of this cloud was then estimated as

\[
(B_{\text{AIRS, CLOUD}} - B_{\text{PAERI, CLOUD}}) - (B_{\text{AIRS, CLEAR}} - B_{\text{PAERI, CLEAR}}) = (B_{\text{AIRS, CLOUD}} - B_{\text{AIRS, CLEAR}}) - (B_{\text{PAERI, CLOUD}} - B_{\text{PAERI, CLEAR}}).
\]

where \(B_{\text{AIRS, CLOUD}}\) and \(B_{\text{AIRS, CLEAR}}\) are simply calculated as the brightness temperatures of the TOA radiance calculated using LBLRTM and LBLDIS for the clear and cloudy cases. \(B_{\text{PAERI, CLEAR}}\) is the brightness temperature...
of the surface as measured by the PAERI. \( BT_{\text{AIRS,CLOUD}} \) is estimated by adding the additional downward longwave flux from the cloud (integrated over the longwave spectrum, then multiplied by \( \pi \)) to the flux emitted by the surface under clear skies and then calculating the brightness temperature of the new, larger flux value.

[35] The effect of the subvisible cloud at the surface dominates that at the top of the atmosphere. The cloud warms the surface under the PAERI by about 0.04 K in each case, afternoon and late evening. However, the effect at TOA, as measured by AIRS, is negligible since the TOA flux is dominated by emission from the surface, not the thin cloud. Therefore the potential effect of a cloud (with an optical depth of 0.003) on the brightness temperature difference (AIRS minus PAERI) is to induce a slight cold bias of a few hundredths of a Kelvin. In other words, a cloud that affects the 835-cm\(^{-1} \) window radiances by 0.1 mW m\(^{-2} \) sr\(^{-1} \) (cm\(^{-1} \))\(^{-1} \) will create a cloud bias of \(-0.04 \) K in the AIRS/PAERI brightness temperature difference. The bias induced by a subvisible cloud can therefore potentially explain about 25 to 30\% of the observed brightness temperature difference between AIRS and the PAERI. Thicker clouds that go undetected will tend to bias the brightness temperature difference even colder.

6. Conclusions

[56] Antarctica is an adequate site for validation of infrared satellite instruments. Radiometric validation performed in Antarctica complements other experiments that use the warmer sea surface as a validation target. The Antarctic Plateau has several advantages over lower-latitude sites. The primary advantage is that the atmospheric correction necessary to convert ground-based radiation measurements to the top-of-the-atmosphere radiances is quite small ([<0.01 K] because of the high elevation of the Plateau and the dry conditions of the overlying atmosphere. In addition, high-latitude sites used for validation have the advantage of being passed over by polar-orbiting satellites multiple times each day. However, this study shows that it is important to use ground-based radiation instruments with a large enough ground footprint to encompass several sastrugi and, thus, average over the significant infrared radiance differences that develop across the snow dunes. These large temperature gradients occur because of differential heating by the sun and, therefore, can be avoided if validation experiments are conducted at night when the sun is below the horizon.

[57] The Atmospheric Infrared Sounder (AIRS) is performing well at low radiance levels, corresponding to brightness temperatures of about 230 to 250 K. The AIRS radiances compare well with TOA calculations that are derived from ground-based measurements made by the Polar Atmospheric Emitted Radiance Interferometer (PAERI), but are slightly cold in the AIRS M-07, M-08, and M-09 detector bands. At the 95\% confidence level, the mean values of the brightness temperature differences are \(-0.12 \pm 0.24 \) K (M-07), \(-0.15 \pm 0.23 \) K (M-08), and \(-0.13 \pm 0.24 \) K (M-09), which validate the AIRS radiances relative to the TOA calculations at Dome C. Subvisible clouds, if present, may account for some of the cold bias.

Although we carefully screened our comparison cases for clouds in the large AIRS field of view, any undetected clouds will tend to bias the brightness temperature comparisons cold.

[58] These validation results are important because many geophysical parameters that are being derived from AIRS data are retrieved from low radiances, such as concentrations of upper tropospheric water vapor and the microphysical properties of high clouds.

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B. Halter, W. L. Roth, and V. P. Walden, Department of Geography, University of Idaho, 375 S. Line Street, Moscow, ID 83844-3021, USA. (vonw@uidaho.edu)

R. S. Stone, Cooperative Institute for Research in Environmental Sciences, University of Colorado, Boulder, CO 80309, USA.