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Statistics Of Arctic Cloud Downwelling Infrared Emissivity

Steven P. Neshyba
University of Puget Sound, nesh@pugetsound.edu

Carsten Rathke
Institut für Weltraumwissenschaften, Freie Universität Berlin, Berlin, Germany

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1. Introduction

[1] Time series of optical depth of Arctic stratus clouds are investigated for scaling properties and biases with respect to a plane-parallel model. The study is based on 3 years of infrared spectrometer and microwave radiometer measurements made at the Atmospheric Radiation Measurement North Slope of Alaska site. Power spectra of radiance and radiance emissivity are found to indicate scaling with a spectral coefficient on the order of 5/3 over 0.5–10 hours, consistent with the Kolmogorov-Obukhov prediction [Kolmogorov, 1941] for three-dimensional turbulence. Irradiance emissivities inferred from the data set, using an independent column approximation and 6-hour time intervals, are further analyzed to find reduction factors for the same mean optical depth in a plane-parallel representation. These factors are estimated to average 0.82 in March and 0.48 in September but with a high degree of variability: The most inhomogeneous quarters of these data sets exhibit reduction factors of 0.57 (March) and 0.20 (September). Observed reduction factors for radiance and irradiance are found to depend primarily on the ratio of mean optical depth to variance, a result consistent with exact results for a $\gamma$ distribution of cloud thickness.

INDEX TERMS: 3320 Atmospheric Composition and Structure: Cloud physics and chemistry; 3250 Mathematical Geophysics: Fractals and multifractals; 3332 Meteorology and Atmospheric Dynamics: Mesospheric dynamics; 3360 Meteorology and Atmospheric Dynamics: Remote sensing; KEYWORDS: Arctic, cloud, inhomogeneity, radiance

in irradiance emissivity between observations and plane-parallel cloud geometry, the latter being a common representation in climate models [Tiedtke, 1996].

2. Optical Depth and the Bispectral Retrieval Algorithm

[5] The connection between vertically integrated optical depth, \( \delta \), and cloud radiance emissivity, \( \varepsilon_{\text{rad}} \), is provided by the effective emissivity relation [Rathke and Fischer, 2002]

\[
\varepsilon_{\text{rad}} \approx 1 - \exp(-k/\cos \theta)
\]

(1)

where \( \theta \) is the angle of view with respect to zenith. Here \( \delta \) and \( \varepsilon_{\text{rad}} \) are wavelength-dependent, as well as functions of time and horizontal coordinates. Equation (1) implicitly accounts for scattering effects and partial cloud cover within a given (narrow) field of view, and therefore, \( \varepsilon_{\text{rad}} \) and \( \delta \) must be interpreted as an “effective” emissivity and optical depth. They depend on cloud microphysical properties (particle phase and size, water content), as well as macrophysical properties (height and thickness). Here \( \delta \), for example, is proportional to the vertically integrated cloud liquid or ice water content, the liquid (or ice) water path (LWP), using the approximation

\[
\delta \approx k_{\text{abs}} \text{LWP}.
\]

(2)

At a wavelength of 962 cm\(^{-1} \), a spherical droplet of liquid water with radius 10 \( \mu \)m is predicted by Mie theory to have an absorption coefficient of \( k_{\text{abs}} = 0.0614 \) m\(^2\)/g [Wiscombe, 1980].

[6] Within a given time window, the radiances corresponding to a zenith angle determine a probability density function for optical depth, \( \rho(\delta) \). Then the time average of radiances corresponding to a zenith angle varies as

\[
\langle \varepsilon_{\text{rad}}(\theta) \rangle = \int_0^\infty \rho(\delta) \left( 1 - e^{-\delta/\cos \theta} \right) d\delta.
\]

(3)

The angle brackets indicate an average over a 6-hour window; that is, we assume ergodicity.

[7] The emissivity associated with downwelling hemisphere-integrated radiances, hereafter “irradiance emissivity,” is obtained next. We associate with any given zenith angle a vertical optical depth of a single column corrected by \( 1/\cos \theta \). This is the “independent column approximation” (ICA) [Ronholm et al., 1980; Cahalan et al., 1994]. The irradiance emissivity is then

\[
\langle \varepsilon_{\text{irrad}} \rangle = \int_0^{\pi/2} 2 \langle \varepsilon_{\text{rad}}(\theta) \rangle \sin \theta \cos \theta \, d\theta,
\]

(4)

where we have assumed \( \langle \varepsilon_{\text{rad}} \rangle \) does not depend on azimuthal angle.

[8] Averages such as appearing in equation (3) may be taken for observed or model distributions of \( \delta \). For example, we define a plane-parallel horizontal (PPH) emissivity average by using a monodispersion for \( \rho(\delta) \), so that \( \langle \varepsilon_{\text{rad}}(\text{PPH}) \rangle \) is given by equation (1) with \( \delta = \langle \delta \rangle \), and

\[
\langle \varepsilon_{\text{rad}}(\text{PPH}) \rangle = 1 - e^{-\langle \delta \rangle} \left( 1 - \langle \delta \rangle \right) - \langle \delta \rangle^2 Ei(\langle \delta \rangle),
\]

(5)

where \( Ei(\langle \delta \rangle) \) is the exponential integral. In the zero-scattering approximation, equation (5) is approximated as

\[
\langle \varepsilon_{\text{rad}}(\text{PPH}) \rangle \approx 1 - e^{-F(\langle \delta \rangle)},
\]

(6)

where \( F \) is the diffusivity factor, 1.66 [Paltridge and Platt, 1976].

[9] Regarding the observational record, a first inclination is to use liquid water path values from microwave radiometer measurements [Liljegren, 1994] to deduce a cloud’s effective emissivity. The ARM data set at Barrow is nearly continuous over the 3 years considered here (1999–2001), recorded at 5-min intervals. Two factors make such an approach incomplete for our purpose, however. First, ice occurs commonly in Arctic clouds, but microwave radiometer retrievals are largely insensitive to ice content. Second, LWP retrievals from microwave radiometry [Liljegren, 1994] exhibit high relative errors when cloud thickness is less than 20 g/m\(^2\) [Dong et al., 2000], whereas such clouds are commonplace in the Arctic [e.g., Curry et al., 1996].

[10] Therefore we present here an algorithm for inferring optical depths based on downwelling infrared radiances measured at the North Slope of Alaska (NSA) ARM site by the Atmospheric Emissivity Radiometer Interferometer (AERI) [Smith et al., 1995; Knuteson et al., 1999]. Figure 1 is presented to show that the influence of clouds in these spectra is most evident in the 8–12 \( \mu \)m (750–1250 cm\(^{-1} \)) window, in particular in “microwindows,” frequencies between lines of strong radiative emission from gases. Figure 1 also demonstrates the sensitivity of the approach, clearly distinguishing a cloud with \( \delta = 0.1 \) from another with \( \delta = 1.5 \) (at 962 cm\(^{-1} \)). These optical depths were inferred using a bispectral retrieval algorithm, which we now describe.

[11] Given a cloud temperature \( T_{\text{cld}} \), \( \delta(\nu) \) gives rise to a cloud downwelling radiance given by

\[
I(\nu) \approx B(\nu, T_{\text{cld}}) \epsilon(\nu),
\]

(7)

received at the ground according to

\[
I(\nu) = t(\nu) \{ B(\nu, T_{\text{cld}}) \epsilon(\nu) + B(\nu, T_{\text{bkg}}) [1 - \epsilon(\nu)] \} + B(\nu, T_{\text{atm}}) \cdot [1 - t(\nu)],
\]

(8)

where \( I \) is the observed downwelling radiance, \( B(\nu, T_{\text{cld}}) \) is the Planck function at \( T_{\text{cld}} \), \( t(\nu) \) is the atmospheric transmissivity from the cloud to the surface, \( T_{\text{atm}} \) is the average temperature of the atmosphere below the cloud, and \( T_{\text{bkg}} \) is the radiative temperature corresponding to the “background,” clear-sky downwelling radiance. (For the remainder of this section “\( \nu \)” refers to wavenumber).

[12] Figure 2 shows time series of downwelling radiance in the 1233 and 962 cm\(^{-1} \) microwindows, expressed in equivalent brightness temperatures. During March 2000, many measurements correspond to clear-sky conditions. These appear as low brightness values, \( \approx 180 \) K in the 1233 cm\(^{-1} \) microwindow and \( \approx 150 \) K in the 962 cm\(^{-1} \) microwindow. Also shown are clear-sky brightness temperatures in these microwindows, based on the subarctic winter model atmosphere of McClatchey et al. [1972]. The approximate agreement (5–10 K root-mean-square error) leads us to the conclusion that “clear-sky” parameters \( t \),
for comparison. confidence because there are fewer cloudless measurements
model atmosphere are satisfactory throughout the month.
frequencies
that the retrieval error for
consider only the cloud contribution in equation (8), i.e.,
equation (7), so that we obtain

and, after rearranging,

where $c_2$ is Planck’s second constant. The retrieval error for $T_{\text{cld}}$ is thus a function of $T_{\text{cld}}$ and $\alpha$, $\nu_1$ and $\nu_2$, the noise $\Delta I_1$, and the effective emissivity ratio $\alpha$ and its standard deviation $\Delta \alpha$:

\[
\Delta T_{\text{cld}} \approx \left( \frac{T_{\text{cld}}}{\ln \left( \frac{I(\nu_1)}{I(\nu_2)} \nu_2^3 \right) I_1} \right)^2 + \left( \frac{T_{\text{cld}}}{\ln \left( \frac{I(\nu_1)}{I(\nu_2)} \nu_2^3 \right) \alpha} \right)^2 .
\]

We have calculated $\alpha$ and $\Delta \alpha$ with the radiative transfer model DISORT [Stamnes et al., 1988] for all microwindows in the 770–2600 cm$^{-1}$ spectral range for the range of cloud/surface temperature contrasts expected in the Arctic spring and summer and for liquid water and ice clouds. The clouds were assumed to be composed of spherical particles, distributed according to a Gamma-Hansen [Hansen, 1971] size distribution, characterized by an effective particle radius $r_{\text{eff}}$ and an effective variance $v_{\text{eff}}$. Narrow and broad size distributions were considered ($v_{\text{eff}} = 0.021$ and $v_{\text{eff}} = 0.370$ for liquid water clouds with $r_{\text{eff}} = 2, 4, \ldots, 16$ μm, and $v_{\text{eff}} = 0.050$ and $v_{\text{eff}} = 0.400$ for ice clouds with $r_{\text{eff}} = 5, 10, \ldots, 40$ μm).

[15] Having $\alpha$ and $\Delta \alpha$, we have estimated $\Delta T_{\text{cld}}$ with equation (11) for all pairs of microwindows in the 770–2600 cm$^{-1}$ spectral range. The smallest error was found for $\nu_1 = 962$ cm$^{-1}$ and $\nu_2 = 1233$ cm$^{-1}$, so that we based our bispectral algorithm on radiance observations in these microwindows. The algorithm consists of two steps: (1) assuming $T_{\text{bkg}}$ and $T_{\text{atm}}$, a primitive atmospheric correction is applied and $I'$ is determined

\[I'(\nu_1) = I(\nu_1) - [1 - t(\nu_1)]B(\nu_1, T_{\text{atm}})\]

\[I'(\nu_2) = I(\nu_2) - [1 - t(\nu_2)]B(\nu_2, T_{\text{atm}})\]

Figure 1. ARM/AERI spectra taken at Barrow, Alaska, on 1 March 2001. The bold line is the spectrum of a cloud with retrieved optical depth of 0.1 at 962 cm$^{-1}$, taken at 2130 GMT. The thin line is the spectrum of a cloud with retrieved optical depth of 1.5 at 962 cm$^{-1}$, taken at 0334 GMT. Radiance units are mW/(m$^{-2}$ St cm$^{-1}$).

Figure 2. Time series of downwelling infrared radiance (expressed in brightness temperatures) measured by the AERI in (a) March 2000 and (b) August 2000 at the NSA ARM site. Brightness temperatures are averaged over the width of the microwindows centered at 962 cm$^{-1}$ (959.998–964.337 cm$^{-1}$) and 1233 cm$^{-1}$ (1230.976–1235.316 cm$^{-1}$). The dotted lines indicate clear-sky $T_{\text{bkg}}$ values calculated for these microwindows and the McClatchey et al. [1972] (a) subarctic winter and (b) subarctic summer atmospheres. See color version of this figure in the HTML.
and (2) assuming \( \alpha, T_{\text{bkg,1}} \) and \( T_{\text{bkg,2}}, T_{\text{cld}} \) and \( \varepsilon(v_1) \) are found so that

\[
\frac{I'(v_1) - B(v_1, T_{\text{bkg,1}})}{B(v_1, T_{\text{cld}}) - B(v_1, T_{\text{bkg,1}})} = \varepsilon(v_1) = \alpha \varepsilon(v_2)
\]

\[
= \alpha \frac{I'(v_2) - B(v_2, T_{\text{bkg,2}})}{B(v_2, T_{\text{cld}}) - B(v_2, T_{\text{bkg,2}})}
\]

with \( 0 < \varepsilon(v_1) \leq 1 \) and \( 200 \, \text{K} < T_{\text{cld}} < T_{\text{surf}} + 30 \, \text{K} \). The only additional information required is the surface temperature \( T_{\text{surf}} \), which is provided in the ARM/AERI data set as ambient air temperature at hatch opening. Any measurement for which the two latter conditions cannot be met is labeled “clear.” The parameters needed for the application of this algorithm are supplied in Table 1.

Previously (equation (11)), we neglected errors associated with the atmospheric correction, that is, errors due to errors in the specification of \( t, T_{\text{bkg}} \) and \( T_{\text{atm}} \). More realistic error estimates were obtained by repeating the retrievals with the bispectral algorithm and varying all parameters between their minimum and maximum expected values (Table 1). The results of this error analysis are shown in Figure 4. The maximum values correspond to the background radiance at the height of 1 km in these atmospheres but with the water vapor column doubled in all layers.

The minimum values correspond to the transmission from the surface to a height of 1 km in these atmospheres but with the water vapor column halved in all layers. The maximum values correspond to the transmission from the surface to the tropopause in these atmospheres but with the water vapor column doubled in all layers.

The minimum values correspond to the average radiative temperature of these atmospheres from the surface to the tropopause. The maximum values correspond to the average radiative temperature of these atmospheres from the surface to a height of 1 km.

3. Results and Discussion

Emissivities and temperatures retrieved for two months using the bispectral algorithm are presented in Figure 5. These two months were selected because of a near-continuous record of the ARM/FTIR data set. Qualitatively, it is clear that retrieved emissivities track the radiance very closely, as expected. For August 2000, the variability in retrieved temperature is quite small in relative terms (smaller than 5% of the mean value for the month) compared to the variability of retrieved optical depths, which spans the full physically permissible range (0–1) on an almost daily basis. For March 2001, the variability of temperature is greater but (in relative terms) still much less than the variability of the emissivity.

Power spectra of the downwelling radiance and emissivity are presented in Figure 6. Straight line fits to octave-averaged points appear to indicate scaling over 0.5–10 hours. With the exception of radiance for March 2001, spectral coefficients are within ±0.1 of the Kolmogorov-Obukhov prediction [Kolmogorov, 1941] for the density fluctuation of a passive scalar in three-dimensional turbulence, 5/3. It is not known whether the unexpectedly high spectral coefficient for radiance in March 2001 (\( \beta_R = 1.93 \pm 0.09 \)) is representative of the month; considering that the corresponding emissivity exhibits a significantly lower value (\( \beta_e = 1.76 \pm 0.06 \)), a possible explanation is that the radiance power spectrum includes scaling in temperature as well as in optical depth.

In order to simulate a ∼200–300 km spatial grid scale of a global climate model, we have broken each monthly data set into 6-hour windows. The windows are advanced in increments of 3 hours, yielding eight values of \( \langle \delta \rangle \) and eight values of the homogeneity parameter \( \nu = (\langle \delta \rangle / \sigma)^2 \) per day. Moreover, windows with \( \langle \delta \rangle \) smaller than 0.2 were discarded, so that the analysis would address only detected cloud. Figure 7 is a scatterplot of the March and September data sets, spanning years 1999–2001, processed in this way. Geometric means of these distributions are displayed in Table 2. Although quite broad, the distributions in \( \langle \delta \rangle \) for the two months are clearly distin-

### Table 1. Parameters for the Bispectral Retrieval Algorithm

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Subarctic Winter Atmosphere</th>
<th>Subarctic Summer Atmosphere</th>
</tr>
</thead>
<tbody>
<tr>
<td>( v_1 )</td>
<td>962 cm(^{-1} )</td>
<td>962 cm(^{-1} )</td>
</tr>
<tr>
<td>( v_2 )</td>
<td>1233 cm(^{-1} )</td>
<td>1233 cm(^{-1} )</td>
</tr>
<tr>
<td>( \alpha^a )</td>
<td>0.98 ± 0.04</td>
<td>0.98 ± 0.04</td>
</tr>
<tr>
<td>( T_{\text{bkg}} )</td>
<td>130 (100–159)</td>
<td>164 (127–181)</td>
</tr>
<tr>
<td>( T_{\text{cld}} )</td>
<td>0.98 (1.00–0.96)</td>
<td>0.96 (0.99–0.93)</td>
</tr>
<tr>
<td>( T_{\text{atm}} )</td>
<td>252 (248–256)</td>
<td>247 (246.8–247.3)</td>
</tr>
</tbody>
</table>

\( ^a \)Average is given plus/minus standard deviation.

\( ^b \)The minimum values correspond to the background radiance at the height of the tropopause (9 km in winter, 10 km in summer). The maximum values correspond to the background radiance at the height of 1 km in these atmospheres but with the water vapor column doubled in all layers.

\( ^c \)The minimum values correspond to the transmission from the surface to a height of 1 km in these atmospheres but with the water vapor column halved in all layers. The maximum values correspond to the transmission from the surface to the tropopause in these atmospheres but with the water vapor column doubled in all layers.

\( ^d \)The minimum values correspond to the average radiative temperature of these atmospheres from the surface to the tropopause. The maximum values correspond to the average radiative temperature of these atmospheres from the surface to a height of 1 km.
guishable from each other, with generally thicker clouds in September. This pattern is consistent with known seasonality of cloudiness in the Arctic (see, e.g., Intrieri et al. [2002b] and Curry et al. [1996] for recent climatologies). Figure 7 also suggests that September and March are not nearly so distinguishable in terms of homogeneity as they are in terms of thickness. As Table 2 shows, the geometric mean of the homogeneity parameter for March, $\nu = 3.3$, is slightly higher than for September, $\nu = 2.4$. For comparison, the “overcast stratocumulus” cloud classification of Barker et al. [1996] type “A,” spans the range $1.5 < \nu < 22$, whereas the “broken stratocumulus classification,” type “B,” spans $0.4 < \nu < 3$ [Barker et al., 1996].

Figure 3. Retrieval error in $T_{clad}$ and $\varepsilon(962 \text{ cm}^{-1})$ as a function of $\varepsilon(962 \text{ cm}^{-1})$ for the bispectral retrieval algorithm for clouds located in the McClatchey et al. [1972] (a) subarctic winter atmosphere and (b) subarctic summer atmosphere. The curves were obtained by varying the different parameters entering equation (8).

The distributions in $\nu$ displayed in Figure 7 are wide enough to include many 6-hour periods for which clouds are much more inhomogeneous, however. Results considering only the most inhomogeneous one fourth of 6-hour measurement periods are displayed in Table 3. These subsets of March and September fall generally inside category “B,” broken stratocumulus.

Figure 4. Optical depths at 962 cm$^{-1}$ retrieved from downwelling infrared radiance data according to the bispectral algorithm (solid line), and microwave data, based on ARM/MWR retrievals of LWP [Liljegren, 1994] (dashed-dotted) during March 2001. The origin is midnight of March 1, 2001 (GMT).
determines the spectral coefficient \( C_{\text{AHalan}}, 1994, \) equation (B2)]. Another distribution function is the \( \gamma \) distribution, \( \gamma(\nu) \equiv \frac{1}{C_{0}} (n) n \cdot d n / C_{18} C_{19} n \cdot d n / C_{0} / \exp[1/2] \) \( \gamma(\nu) \equiv \frac{1}{C_{0}} (n) n \cdot d n / C_{18} C_{19} n \cdot d n / C_{0} \) \( \exp[1/2] \) \( B_{\text{arker}}, 1996 \). \( C_{\text{AHalan}} \)'s zeroth-order approach to the \( \gamma \) distribution also produces the functionality \( \chi_{\text{rad}} = f(\nu), \)

\[
\rho(\delta) = \frac{1}{\Gamma(\nu)} \left( \frac{\nu}{\delta} \right)^{\nu-1} \exp\left(-\frac{\nu}{\delta}\right) \tag{16}
\]

\[
\chi_{\text{rad},\gamma0} = \prod_{\nu=0}^{\infty} \left( 1 + \frac{1}{\nu + k} \right) \exp\left(-\frac{1}{\nu + k}\right) \tag{17}
\]

\( B_{\text{arker}}, 1996 \).

Figure 5. Recorded radiance, retrieved emissivity, and retrieved cloud temperature for selected months in 1998–2001. The origin in each graph is midnight GMT of the first of the month. Emissivities correspond to a frequency of 962 cm\(^{-1}\).

Figure 6. Power spectra of the radiance and emissivity appearing in Figure 5. Straight lines are fit to octave-averaged data points. Uncertainties are at the 95% confidence level based on deviations of these points from the best fit straight line shown.
Exact relationships, not using the zeroth-order approach, are obtainable using numerical and analytical methods. For example, numerical realizations of the bounded cascade model (BCM) in optical depth lead to $\chi_{\text{rad}}^{\text{BCM}}$ using equation (3) and to $\chi_{\text{rad},\text{BCM}}$ using equation (14). Alternatively, analysis of the $\gamma$ distribution of equation (16) leads to the exact result,

$$\chi_{\text{rad},\gamma} = \frac{\nu \cos \theta}{\langle \delta \rangle} \log \left( 1 + \frac{\langle \delta \rangle}{\nu \cos \theta} \right). \quad (18)$$

That is, unlike the zeroth-order prediction, functionality of the form $\chi_{\text{rad}} = \chi_{\text{rad}}(\nu \cos \theta/\langle \delta \rangle)$ is suggested. Consequently, we have chosen to display in Figure 8 reduction factors as a function of the ratio $\nu/\langle \delta \rangle$ $(\cos \theta = 1$ for these measurements). Also shown are exact analytical results for a $\gamma$ distribution, equation (18), and exact numerical results for the bounded cascade model, obtained as described above. Although there is considerable scatter, the trend is clearly one that favors the $\gamma$ distribution, for which the root-mean-square error in $\chi$ is 0.07. We also display reduction factors based on the zeroth-order approximation for the $\gamma$ distribution, equation (17); this function is graphed assuming $\langle \delta \rangle = 4$, but the overall match to observations is no better at other values. Using the ratio of geometric-mean values of $\nu$ and $\langle \delta \rangle$ from Tables 2 and 3, equation (18) allows us to infer corresponding reduction factors, also presented in Tables 2 and 3.

Using the zero-scattering approximation (equation (6)) in equation (15), the reduction factor for irradiance also appears as an exclusive function of $\nu/\langle \delta \rangle$,

$$\chi_{\text{irrad},\gamma} \approx \frac{\nu}{F(\langle \delta \rangle)} \log \left( 1 + \frac{F(\langle \delta \rangle)}{\nu} \right). \quad (19)$$

with $\gamma$ function and zero-scattering approximated. This function is displayed in Figure 9 along with observations. Observations deviate from equation (19) with a root-mean-square error of 0.08.

4. Conclusions

We have analyzed time series of Arctic cloud optical depth, determined from AERI and microwave radiometer

<table>
<thead>
<tr>
<th>Year</th>
<th>$\nu^a$</th>
<th>$\langle \delta \rangle^b$</th>
<th>$\chi_{\text{rad}}^c$</th>
<th>$\chi_{\text{rad}}^d$</th>
<th>$\varepsilon_{\text{rad}}^e$</th>
<th>$\varepsilon_{\text{rad}}^f$</th>
<th>$\chi_{\text{irrad}}^g$</th>
<th>$\varepsilon_{\text{irrad}}^h$</th>
</tr>
</thead>
<tbody>
<tr>
<td>1999</td>
<td>September</td>
<td>2.4</td>
<td>4.0</td>
<td>0.59</td>
<td>8</td>
<td>0.48</td>
<td>4</td>
<td>9</td>
</tr>
<tr>
<td>March</td>
<td>3.3</td>
<td>0.9</td>
<td>0.88</td>
<td>8</td>
<td>0.82</td>
<td>6</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

$^a$Value is geometric mean of 6-hour homogeneity values ($\nu$).
$^b$Value is geometric mean of 6-hour mean optical depth values ($\langle \delta \rangle$).
$^c$Value is based on the ratio of geometric means ($\nu/\langle \delta \rangle$) and equation (18).
$^d$Value is based on the ratio of geometric means ($\nu/\langle \delta \rangle$) and equation (19).

<table>
<thead>
<tr>
<th>Year</th>
<th>$\nu^a$</th>
<th>$\langle \delta \rangle^b$</th>
<th>$\chi_{\text{rad}}^c$</th>
<th>$\chi_{\text{rad}}^d$</th>
<th>$\varepsilon_{\text{rad}}^e$</th>
<th>$\varepsilon_{\text{rad}}^f$</th>
<th>$\chi_{\text{irrad}}^g$</th>
<th>$\varepsilon_{\text{irrad}}^h$</th>
</tr>
</thead>
<tbody>
<tr>
<td>1999</td>
<td>September</td>
<td>1.1</td>
<td>8.9</td>
<td>0.27</td>
<td>10</td>
<td>0.20</td>
<td>4</td>
<td>11</td>
</tr>
<tr>
<td>March</td>
<td>1.4</td>
<td>1.5</td>
<td>0.68</td>
<td>22</td>
<td>0.57</td>
<td>13</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

$^a$Value is geometric mean of 6-hour homogeneity values ($\nu$).
$^b$Value is geometric mean of 6-hour mean optical depth values ($\langle \delta \rangle$).
$^c$Value is based on the ratio of geometric means ($\nu/\langle \delta \rangle$) and equation (18).
$^d$Value is based on the ratio of geometric means ($\nu/\langle \delta \rangle$) and equation (19).
measurements made at the ARM NSA site, in order to
to investigate the impact of horizontal cloud inhomogeneity on
downward longwave radiation, an important variable for
to the cold climate. The horizontal variability of cloud thickness
retrieved from the measurements leads to irradiance plane-
parallel reduction factors ($\chi_{\text{irrad}}$) of 0.48 typical of Septem-
ber, and 0.82 typical of March, based on 6-hour time
windows, using the independent column approximation.
The range of reduction factors is quite large, with the most
horizontally inhomogeneous quarter exhibiting reduction
factors as low $\chi_{\text{irrad}} \sim 0.2$ (for September). Observed
reduction factors, also calculated within the ICA, compare
favorably with exact values based on the $\gamma$ distribution, with
root-mean-square deviations of 0.07–0.08.

[27] The corresponding irradiance emissivity biases are
on the order 4–13%. For comparison with actual irradiance
values, Intrieri et al. [2002a] showed that the net surface
cloud radiative forcing at the SHEBA ice camp during the
winter of 1998 averaged about 20 W m$^{-2}$ during that time,
with an uncertainty of 15%. During a 3-day period in April
1998 reported by Intrieri et al. [2002a], the downwelling
longwave irradiance attributable to clouds reached a maxi-
mum of 60 W m$^{-2}$, which is about 25% of the total
longwave downward irradiance. Thus biases of 13% in this
cloud forcing should not be neglected.

[28] Considering cloud optical depth as a proxy for cloud
thickness, these results argue for inclusion in climate
models of the radiative effects of cloud horizontal variabil-
ity in a way that depends on the ratio $\langle h \rangle / (\langle h \rangle / \sigma^2)$. Presently, climate models either ignore subgrid
cloud variability or use a constant reduction factor for all
types of clouds [Tiedtke, 1996]. If a climate model were to
predict the variance ($\sigma^2$) as well as mean ($\langle h \rangle$) of cloud
thickness for a given grid cell, Figure 9 could provide a
useful indication of the irradiance reduction factor likely to
occur, as well as the uncertainty in that value.

[29] We should qualify these conclusions by restating
approximations or assumptions that might be relevant. First,
estimates of the plane-parallel bias made in section 3 are
based on an ergodic assumption, that statistics in time are
the same as statistics in space. To address this objection,
one would need ground-based spatiotemporal information
not available in the ARM/AERI/MWR archive for Barrow.
Second, the ICA is applied throughout. Finally, considering
that cloud temperature usually shows much less horizontal
variability than cloud optical depth [Rathke et al., 2002]
(see also Figure 5), it can be expected that these results will
not be affected greatly by inclusion of temperature variabil-
ity. However, we have not undertaken a quantitative exam-
ination of this statement.

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Figure 8. Reduction factors for irradiance emissivity, $\chi_{\text{irrad}}$.

Figure 9. Reduction factors for irradiance emissivity, $\chi_{\text{irrad}}$. 

\begin{figure}[h]
\centering
\includegraphics[width=\textwidth]{figure8.png}
\caption{Reduction factors for irradiance emissivity, $\chi_{\text{irrad}}$.}
\end{figure}


