Reflected Fluxes for Broken Clouds over a Lambertian Surface

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ABSTRACT

Reflected fluxes are calculated for broken cloudiness (i.e., nonplane parallel) as a function of cloud cover, cloud optical depth, solar zenith angle and surface albedo. These calculations extend previous results for broken cloud reflected fluxes over a black surface. The present study demonstrates that not only radiances but also radiative fluxes over high albedo surfaces may be decreased by the presence of broken cloudiness. Conventional wisdom states that cloud radiances (brightnesses) are always greater than the background. While most cloud retrieval schemes are built around this assumption, it is incorrect for clouds over high albedo surfaces such as found in polar regions. However, the most startling and counterintuitive conclusion of this study is that nonabsorbing finite clouds over a highly reflecting surface will decrease the system albedo. As a result, surface absorption is increased, the result of multiple scattering between surface and cloud layer, controlled by cloud morphology and cloud optical thickness. A simple parameterization of the effects of cloud contamination upon retrieved albedo is given in terms of solar zenith angle, cloud optical depth, surface albedo, cloud cover, and plane-parallel cloud albedo. In this way, the effects of broken cloudiness are modeled in terms of easily computed plane-parallel values.

1. Introduction

The main objectives of the World Climate Research Program are to monitor changes in global climate and to gain an understanding of probable climate responses to various perturbations of the environment. As part of this effort, the International Satellite Cloud Climatology Project is focused upon achieving an accurate, global cloud climatology (Schiffer and Rossow 1983). Clouds are the greatest attenuation of radiation in the atmosphere and the net radiative balance is globally dominated by the cloud albedo effect (Herman et al. 1980; Hartmann and Short 1980). However, there remain validation problems for global cloud distributions between the various cloud datasets.

In a similar effort, the International Satellite Land Surface Climate Project (ISLSCP) is focused on the remote sensing of land-surface processes. Important variables at the surface include: albedo, incident solar radiation, the partition of energy, surface temperature, soil moisture, evapotranspiration, and vegetation. When present, clouds are the major determinants of radiative flux at the ground (Pinker and Ewing 1985; Stowe 1984; Justus et al. 1986). Errors in derived surface variables such as sensible heat flux due to cloudiness can be very large (Camuffo and Bernardi 1982). Abdellaoui et al. (1986) find that a 30% change in incident solar radiative flux at the ground can lead to variation in thermal inertia and evapotranspiration by about a factor of two.

Studies of the climatic impact of increasing atmospheric trace gases suggest that global greenhouse warming will soon rise above the level of natural climate variability (e.g., Hansen et al. 1988). By the early 1990s, there should be a noticeable increase in global surface temperatures. This warming is expected to be most apparent in the Arctic Ocean and in ocean areas near Antarctica, resulting in greater melting of snow and sea ice. Because the darker melted surfaces cause a lowering of regional albedo and greater absorption of solar radiation, the warming is expected to be amplified by factors of 1.5–3 in the polar regions (Wetherald and Manabe 1986; Schlesinger and Mitchell 1987; Steffen and Lewis 1988).

One aspect of a warmer global climate is expected to be a more active hydrological cycle. Warmer land and ocean surfaces lead to greater evaporation. Because the atmosphere tends to conserve relative humidity with change in temperature (Manabe and Wetherald 1967), a higher absolute humidity is expected to result. Increased atmospheric humidity can be expected to
alter global cloud distributions (Cess 1976), although the nature of these changes is unclear. The largest uncertainty in prediction of climate impacts on greenhouse forcings is in predicting how clouds will respond to and alter radiative fluxes (Ramanathan 1988).

Arctic stratus has been shown to be very important to the arctic heat balance (Vowinckel and Orvig 1962), directly affecting surface melt (Parkinson et al. 1987). Because increased cloud cover is associated with ice breakup and increased open water (Robinson et al. 1986), one expected observable trend for global warming is increased polar cloud cover, especially in the Arctic Ocean. Polar cloud cover changes and their attendant changes in radiative fluxes are expected to have a significant effect on sea ice conditions (Shine and Crane 1984) and on regional ice-albedo feedback (Barry et al. 1984). However, our present understanding of the effects of broken cloudiness over various surfaces is insufficient for the development of climate model parameterizations.

The present study examines the effect of broken clouds over surfaces of varying surface albedo on solar reflected fluxes at the top of the atmosphere. Section 2 examines the radiances from clouds and ice/snow surfaces in polar regions, showing that clouds often are darker than the underlying surface. Section 3 describes the method of Monte Carlo simulations of cloud-surface albedo and section 4 presents the results. Section 5 presents a method to parameterize the cloud-background system albedo and section 6 discusses the significance of the results.

2. Cloud and surface radiances in polar regions

Numerous studies have been undertaken to retrieve surface albedos from satellite observations and to discriminate cloud cover. The standard assumption in most of these algorithms is that cloud cover increases the observed albedo (e.g., Moser and Raschke 1984). Therefore, a threshold level is set for cloud detection (e.g., Reynolds and Vonder Haar 1977; Minnis and Harrison 1984; Wielicki et al. 1986). However, considerable errors occur in this process due to pixels which are only partially filled (Wielicki and Welch 1986; Minnis and Wielicki 1988). Also, albedo differences alone are generally insufficient to distinguish reliably between different cloud types and surface features, especially over polar regions. This is due to the similarity in cloud and ice spectral signatures in both visible and infrared wavelengths (Barry 1983). Examples are shown in Fig. 1. Here each of the six LANDSAT MSS scenes show Arctic or Antarctic regions with partial cloud cover. Each scene is 185 km wide by 170 km long with spatial resolution of 57 m. First, note that cloud-covered regions have radiances approximately equal to, or even lower than, the underlying ice-covered surfaces. In such cases, visible channel threshold techniques are inapplicable. Cloud radiance is lower than ice radiance, in part due to the differences in bidirectional reflectance functions (Suttle et al. 1981; Stuhlmann et al. 1983). It is not true, as is tacitly assumed in virtually all cloud climatology retrieval schemes, that clouds are always brighter than the underlying surface.

The very dark regions in Fig. 1 are the ocean surface and cloud shadows. Since the solar elevation is known, cloud shadows allow cloud base height to be determined, a very important quantity affecting climate energy balance. After all, climate equilibrium is, in its simplest terms, the balance between solar heating and infrared cooling. Cloud height strongly influences local infrared cooling. The gray, noncloud-covered regions in Fig. 1 are drainage canals, vertical melt holes, and scattered melt ponds (Hall and Martinic 1985). After melting reduces ice thickness to 1–1.5 m, the ice tends to crack along the many drainage-canal flows, aided by the action of tides and storms (Washington and Parkinson 1987).

Of note in Fig. 1 are the small scale structures inherent to the low level stratocumulus cloudiness. Such structures also are observed in fogs (Welch and Wielicki 1986) and in marine stratocumulus off the coast of California (Welch et al. 1988). Herman (1977) reported that polar stratocumulus are horizontally inhomogeneous with small scale variations (≤5 km). The cloud tops frequently are characterized by 10–30 m turrets, while bases often are found with 10–20 m “scud” elements protruding from cloud bottom. Small scale structure strongly impacts cloud radiative fluxes (Welch and Wielicki 1984; Foot 1988) which, in turn, affects climate assessment derived from remotely sensed radiances. The climatically important radiative fluxes then are derived from remotely sensed radiances using bidirectional reflectance functions (e.g., the Earth Radiation Budget Experiment).

Multispectral signatures are used in most cloud detection algorithms (Greaves and Chang 1970; Shenk and Solomonson 1972; Shenk et al. 1976; Reynolds and Vonder Haar 1977; Parikh 1977; Minnis and Harrison 1984; Saunders and Kriebel 1988; McGuffie et al. 1988). In general, it is found that thin cirrus and low clouds are the most difficult to distinguish and classify. During the recent clouds in Climate II Workshop, Rosswow (1987) and others noted that cloud detection remains a serious problem in polar regions using standard classification algorithms. Even with the most sophisticated algorithms, snow covered surfaces tend to be incorrectly assigned as cloudy (Saunders and Kriebel 1988).

Because it is difficult to detect cloud cover over high albedo surfaces and because cloud inhomogeneities have such a large impact on radiative fluxes, it is appropriate to make theoretical assessments of these cases. The present study makes quantitative estimates of the effect of partial cloud cover over surfaces of different albedo using Monte Carlo simulations.
3. Method of calculation

Early attempts to define the properties of broken cloudiness relied upon calculations of the radiative characteristics of individual cloud elements (Busygin et al. 1973; McKee and Cox 1974). However, the proximity of neighboring clouds significantly alters cloud field radiances through cloud–cloud interactions and shading. Investigations of cloud field radiative interactions based upon clouds of identical size and shape have been reported by Aida (1976, 1977), Davies (1978), Gube et al. (1980), Welch and Zdunkowski (1981a,b), Bradley (1981), Harshvardhan (1982), Weinman and Harshvardhan (1982), Claussen (1982), Harshvardhan and Thomas (1984), and Welch and Wielicki (1984, 1986). Cloud shape, mutual shadowing, cloud–cloud interactions, spatial inhomogeneities and surface albedo have been shown to be important variables. Aida (1977) was the first to include cloud size distributions in model calculations. These studies recently have been extended by Kite (1987) and by Cretel et al. (1988). Each of the foregoing simulation studies applies the Monte Carlo method.

The Monte Carlo method assumes that the interaction of a photon with cloud droplets and the ground is independent of any previous interactions. Photon packets are traced through the atmosphere and cloud, accounting for both absorption and scattering. Details concerning the Monte Carlo procedure are given in the aforementioned references and by Marchuk et al. (1980).

The present study assumes clouds to be of cubic shape, 1 km in diameter \( D \), base height of 1 km, with uniform optical properties throughout. Single scattering albedo is taken as \( \omega_0 = 1.0 \) and the phase function is approximated by the Henyey–Greenstein function with asymmetry factor \( g = 0.86 \). Clouds in the field are assumed to be arrayed in regular patterns of horizontally infinite extent (Fig. 2a). While both cloud shape and cloud size distribution alter the resulting predicted fluxes, the overall pattern of variation of fluxes with solar zenith angle and cloud amount is preserved (Aida 1977; Welch and Wielicki 1984; Cretel et al. 1988).

Therefore, these assumptions are deemed adequate for this pilot study of the effect of surface albedo upon radiative fluxes.

Spacing between cloud centers \( (S) \) is varied to simulate cloud cover \( (N) \),

\[
N = (D/S)^2.
\]  

(1)

Mutual shadowing (Fig. 2b) occurs when cloud spacing in the plane of the sun is

\[
S < D + H \tan \theta,
\]  

(2)

where \( H \) is cloud vertical thickness and \( \theta \) is solar zenith angle. Only the top and part of one cloud side are illuminated.

![Fig. 2. (A) Assumed cloud field array composed of cubic shaped clouds of diameter \( D \) with cloud spacing \( S \). Paths of three photons are shown which interact with neighboring clouds; primes indicate the reinsertion points for these photons in the original cloud. (B) Shading by neighboring clouds. (C) Possible photon paths, explained in the text.](image)

Potential photon paths are illustrated in Fig. 2c. Photon packets may enter cloud top \( (1) \) or the cloud side or may propagate directly to the ground \( (2) \).
Within the cloud, photons are scattered by water droplets and absorbed by water droplets and water vapor. Exiting photon packets (A) may be traced to the top of the atmosphere, exiting the cloud top (3) or cloud sides, (B) may enter a neighboring cloud (4), or (C) may propagate to the ground through cloud base (5) or cloud sides. Photon packets at the ground (2 and 5) experience Lambertian reflection with $\alpha$ of the incident energy at the ground being reflected and $(1 - \alpha)$ absorbed. This reflected energy may be traced to the top of the atmosphere (6), back into the original cloud (7), or into a neighboring cloud (8). Multiple scatterings between clouds and ground are allowed. The photon packet is reentered into the original cloud at the equivalent position and angle, as shown by primes in Fig. 2a (i.e., periodic boundary conditions). The photon packet is traced until it is totally absorbed or until it reaches the top of the atmosphere. In contrast to previous studies, the present model

1) does not assume that radiation exiting cloud sides is isotropic;
2) does not neglect multiple scattering between clouds;
3) does include shading by neighboring clouds; and
4) does allow for multiple reflections between clouds and ground. However, cloud shape and size distribution contributions are not included in the present investigation.

4. Cloud and surface albedos

In the following results, optical thickness is taken as $t = 3, 5, 10, 20$ and 50, and calculations are made for solar zenith angles of $\theta = 0^\circ$ and $60^\circ$. Figure 3 shows total albedo as a function of cloud cover and surface albedo. The values at cloud cover $N = 1.0$ are the plane-parallel albedos and are in agreement with calculations based on the delta-Eddington and adding/doubling methods. The values at $N = 0.0$ are albedos of the surface without cloud cover.

This choice of surface albedos covers a wide range of potential conditions. Oceans and large, deep lakes have albedos on the order of 5%; rural agricultural regions have typical albedos of 10%-15%, depending on vegetation cover and soil moisture; urban regions have typical albedos on the order of 20%-25%; deserts have albedos on the order of 30%-35%; and snow and ice have albedos on the order of 40%-80%.

In the following discussion, we use a 5% change in system albedo merely as a small reference value. For a surface albedo of $a = 0.0$ at solar zenith angle $\theta = 0^\circ$, Fig. 3 shows that cloud cover increases the retrieved albedo by 0.05 (5%) for cloud cover $N = 30$% with optically thin clouds of $t = 5$. This same increase in albedo occurs with cloud cover of only $N = 12$% for optically thicker clouds of $t = 20$. Note that the optical thickness of small cumulus typically ranges from about $t = 5$ to about $t = 20$. Cloud cover increases the retrieved albedo from 0.10 to 0.15 at cloud cover $N = 45$% for optically thin clouds ($t = 5$) and at $N = 20$% for optically thick clouds ($t = 20$). As surface albedo increases, progressively larger cloud covers are necessary to increase the system albedo by a set amount.

Note that for surface albedos greater than $a \geq 0.2$, clouds may even decrease the retrieved system albedo value. At $a = 0.2$, optically thin clouds with $t \leq 5$ decrease the retrieved albedo below 0.2 for cloud covers $N \leq 30$%. For $a = 0.5$, even optically thick clouds decrease the retrieval albedo below 0.5 for cloud covers of $N \leq 20$%-30%.

Clearly, the effects of cloud cover are greatest over regions of low surface albedo, such as oceans and agricultural regions. This is because even small cloud covers strongly impact on the system albedo in these regions. Also, rural and agricultural regions are the primary focus of many remote sensing and climate monitoring efforts. In these regions, cloud covers of $N = 10$%-25% can be expected to change the retrieved albedo on the order of $\leq 0.05$.

These variations in albedo increase significantly with increasing solar zenith angle. At $\theta = 60^\circ$ and for surface albedos of $a = 0.0$ and $a = 0.1$, retrieved albedos increase by 0.05 for cloud covers of only $N = 5$%-10%. Therefore, over rural and water surfaces, even clouds of low optical thickness and small cloud cover significantly change the retrieved system albedos. Even over deserts ($a \approx 0.3$), cloud cover of 10%-20% can lead to significant (0.05) changes in albedo at $\theta = 60^\circ$. However, over ice and snow, cloud cover does not change the retrieved albedo by more than 5% until $N > 20$% (at $t = 20$) and $N > 40$% (at $t = 5$).

In order to understand better these processes, it is useful to decompose the various contributions. Table 1 shows the flux components for a representative case of optical thickness $t = 10$, solar zenith angle of $0^\circ$ and $N = 20$% cloud cover, as a function of surface albedo $a$. All values in Table 1 are given as percentages of incoming solar flux. Because the incident photon packets have been entered in a regular grid array for calculations, the exact value of cloud cover is $N = 19.36$% rather than $N = 20$% in Table 1. In the first column, $a$ is the surface albedo. The second column gives the initial downward flux to the ground (Initial Ground 4). This contribution includes radiation propagating directly to the ground $(1 - N)$ plus the radiation scattered downward from the clouds. For this cloud field, approximately 30% of the radiation incident on clouds is reflected back to space; this is 5.81% of the total radiation $(\sim 30\% \times N)$, as entered in column 7. Note that columns 2 and 7 represent initial conditions before surface reflections are taken into account and are independent of surface albedo $a$. The third column shows the initial absorption of incoming radiation at the ground [=$(1 - a)$ of column 2] and the fourth
Fig. 3. Cloud cover albedo estimates as a function of cloud cover $N$, surface albedo $a$, solar zenith angle $\theta$, and cloud optical depth $t$. 
column shows the initial radiation reflected by the surface \((=(a) \text{ of column } 2)\). Radiation at the surface is reflected Lambertian (column 4), approximately 35% of which propagates directly to space; the remainder (or 65% of column 4) enters clouds which re-reflect about 30% of this radiation back to the ground. There are approximately two multiple reflections between ground and clouds before the radiation either is lost to space or is absorbed by the surface. The values of multiply scattered radiation absorbed by the ground are given in column 5, and the total absorbed values by the ground are given in column 6. Note that for a surface albedo of \(a = 0.5\), 52.18% of the initial radiation is finally absorbed by the ground. This is 2.18% more than would have been absorbed without the presence of clouds. The total radiative flux at the top of the atmosphere is 47.83% (column 9), which is less than that from a cloud free surface. Therefore, it is the multiple scattering between surface and clouds (over high albedo surfaces) which leads to a decrease in outgoing radiative fluxes.

Because this result is so counterintuitive, let us examine it in more detail. Table 2 shows radiative fluxes computed for 1) 35% of the radiation reflected from the surface propagating directly to space without entering a cloud and 2) 70% of radiation entering a cloud from below scattered in the upward direction to space. Of the radiation scattered through the cloud in the upward direction, less than half exits through the cloud top. The radiation is virtually depleted after multiple scattering twice between surface and clouds. Clearly, contributions by multiple scattering lead to the decrease in system albedo below that of the background.

These results depend upon both cloud optical thickness and cloud morphology. Table 3 shows the same analysis as Table 1 for optical thickness of \(t = 5, 10, 20\) and 50 for the case of surface albedo \(a = 0.5\). With increasing cloud optical thickness, a smaller amount of total radiation reaches the surface (column 2). However, the optically thicker clouds re-reflect a larger portion of exiting radiation back to the surface. The net effect is that these two contributions tend to cancel, with only a small decrease in total surface absorption for the different cloud optical thicknesses at cloud cover of \(N = 0.2\).

A cloud cover of \(N = 0.5\) has a much larger effect. Comparison of Table 3 (\(N = 0.2\)) with Table 4 (\(N = 0.5\)) for a given cloud optical thickness shows that radiation at the ground (column 2) is decreased both by greater cloud cover and by increased cloud–cloud interactions. For example, for \(t = 10\) and \(\theta = 0^\circ\), the initial cloud reflectance at the top of the atmosphere (column 7) is 5.81% for \(N = 0.1936\) (Table 3) compared to 16.51% for \(N = 0.5\) (Table 4). Scaling by cloud cover alone would provide a value of initial cloud reflectance \(5.81\times0.5/0.1936 = 15\%\) at \(N = 0.5\). The other 1.51% is due to increased cloud–cloud interactions. Table 4 shows that total absorption at the ground (column 6) decreases rapidly as cloud cover \(N\) increases, especially for the thicker clouds; less radiation is reaching the surface.

Finally, consider the impact of cloud aspect ratio and cloud base height. All other parameters held con-
stant. An increase in cloud base height leads to a smaller exit angle for radiation reflected at the ground propagating directly to space. Therefore, a higher cloud base height leads to greater probability of radiation at the surface being multiply scattered back to the surface. Likewise, for cloud optical thickness held constant, an increase in cloud aspect ratio (height/width) leads to increased surface absorption. For example, a cubic cloud field with aspect ratio = 1, optical thickness $t = 20$, and cloud cover of $N = 0.1936$ at $\theta = 0^\circ$ produces surface absorption of 51.26% and radiative flux at the top of the atmosphere of 48.74% (Table 3). Cloud fields composed of equal optical thickness $t = 20$, but with an aspect ratio of $h/w = 2$ have larger surface absorptions of 52.99% and lower reflected fluxes of 47.01%. The dominant effect here is the much greater probability of multiple scattering between surface and clouds for high aspect ratio clouds.

It should be noted that the present calculations assume cubic clouds to be arrayed in regular arrays (Fig. 1a). A cloud shape such as cylinders or hemispheres (Welch and Wielicki 1984) can be expected to lower the surface absorption slightly. The probability of exiting radiation entering a cloud and being reflected back to the surface is lower in these cases. That is, the gaps between clouds (Welch and Wielicki 1984) allow a greater percentage of radiation reflected from the surface to propagate directly back to space. A more realistic hexagonal cloud field pattern, as shown in Fig. 3b of Welch and Wielicki (1984), also increases the probability of reflected surface radiation entering a cloud. Elimination of the long corridors between clouds can lead only to increased total surface absorption and decreased system albedo by increasing multiple scattering between surface and clouds.

### Table 3. Partition of radiative fluxes (%) for the case of $a = 0.5, \theta = 0^\circ$, $N = 0.1936$, as a function of cloud optical thickness $t$.

<table>
<thead>
<tr>
<th>$t$</th>
<th>2 Initial ground</th>
<th>3 Initial absorption</th>
<th>4 Initial ground†</th>
<th>5 Multiple-scattered absorption</th>
<th>6 Total absorption</th>
<th>7 Initial cloud top-atm†</th>
<th>8 Multiple-scattered top-atm†</th>
<th>9 Total top-atm†</th>
<th>10 Total clouds†</th>
</tr>
</thead>
<tbody>
<tr>
<td>5</td>
<td>96.56</td>
<td>48.28</td>
<td>48.28</td>
<td>3.86</td>
<td>52.14</td>
<td>3.43</td>
<td>44.43</td>
<td>47.86</td>
<td>50</td>
</tr>
<tr>
<td>10</td>
<td>94.19</td>
<td>47.10</td>
<td>47.10</td>
<td>5.08</td>
<td>52.18</td>
<td>5.81</td>
<td>42.02</td>
<td>47.83</td>
<td>50</td>
</tr>
<tr>
<td>20</td>
<td>90.72</td>
<td>45.36</td>
<td>45.36</td>
<td>5.90</td>
<td>51.26</td>
<td>9.28</td>
<td>39.46</td>
<td>48.74</td>
<td>50</td>
</tr>
<tr>
<td>50</td>
<td>86.84</td>
<td>43.42</td>
<td>43.42</td>
<td>7.40</td>
<td>50.82</td>
<td>13.15</td>
<td>36.03</td>
<td>49.18</td>
<td>50</td>
</tr>
</tbody>
</table>

### Table 4. Partition of radiative fluxes (%) for the case of $a = 0.5, \theta = 0^\circ$, $N = 0.5\%$, as a function of cloud optical thickness $t$.

<table>
<thead>
<tr>
<th>$t$</th>
<th>2 Initial ground</th>
<th>3 Initial absorption</th>
<th>4 Initial ground†</th>
<th>5 Multiple-scattered absorption</th>
<th>6 Total absorption</th>
<th>7 Initial cloud top-atm†</th>
<th>8 Multiple-scattered top-atm†</th>
<th>9 Total top-atm†</th>
<th>10 Total clouds†</th>
</tr>
</thead>
<tbody>
<tr>
<td>5</td>
<td>90.95</td>
<td>45.48</td>
<td>45.48</td>
<td>7.48</td>
<td>52.96</td>
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<td>10</td>
<td>83.49</td>
<td>41.74</td>
<td>41.74</td>
<td>9.44</td>
<td>51.18</td>
<td>16.51</td>
<td>32.31</td>
<td>48.82</td>
<td>50</td>
</tr>
<tr>
<td>20</td>
<td>74.19</td>
<td>37.10</td>
<td>37.10</td>
<td>10.67</td>
<td>47.77</td>
<td>25.79</td>
<td>26.44</td>
<td>52.23</td>
<td>50</td>
</tr>
<tr>
<td>50</td>
<td>63.81</td>
<td>31.91</td>
<td>31.91</td>
<td>11.28</td>
<td>43.19</td>
<td>36.17</td>
<td>20.64</td>
<td>56.81</td>
<td>50</td>
</tr>
</tbody>
</table>

### 5. Parameterization

In order to be incorporated into general circulation and climate models, or to be used in satellite retrieval algorithms of surface properties, the effects of cloud cover and surface albedo need to be expressed in some easily parameterized form. Usually, it is desirable to express cloud field properties in terms of the better understood and more easily calculated plane-parallel values.

In the presence of broken cloudiness, the retrieved albedo, $a_{cf}$, consists of a mixture of contributions from both clouds and ground. In the plane-parallel approximation, this may be expressed as

$$a_{cf} = a_c N + (1 - N) a,$$  

(3)

where $a_c$ is the plane-parallel cloud albedo, $a$ is the clear sky albedo, and $N$ the cloud cover. This expression [Eq. (3)] has the correct limiting values of $a_{cf} = a$ at $N = 0$ and $a_{cf} = a_c$ at $N = 1$.

Welch and Wielicki (1985) have parameterized cloud field albedo over a nonreflecting surface. Modifying those empirical relationships for reflecting surfaces, we obtain

$$a_{cf} = N' a_c + (1 - N') a,$$  

(4)

where

$$N' = \left\{ \begin{array}{ll}
1 + f H/D \tan \theta, & N \leq N_s \\
(1 - f) + f N'^{-1/2}, & N > N_s
\end{array} \right. - C_{\mu_0} (1 - N^2).$$  

(5)

Shading occurs for cloud cover $N > N_s$, where

$$N_s = \left(1 + \frac{H}{D} \tan \theta\right)^{-2}.$$  

(6)
The correction factor $C$ is expressed, in terms of optical depth $t$, as

$$C = 0.25 - 0.13t^{-2} - 0.05(t/50)^2. \quad (7)$$

The ratio, $f$, of energy exiting the cloud sides in the upward direction is given by

$$f = \frac{\ln[1 + 0.045(\ln t)] + 0.05(\mu_0 - 2) - 1 \times [1 - (t/70)^2] - 0.0015\mu_0 - 0.25(\mu_0 - \mu^*)(t/50)^2(1 + N^2)}{(1 + N^2)}, \quad (8)$$

where $\mu_0 = \cos \theta$ and $\mu^* = \cos 72.5^\circ$. Finally,

$$N^\nu = \frac{N}{2} \left[ 1 + \left( \frac{1 + (H/D) \tan \theta}{N^{-1/2}}, \quad N \leq N_\nu \right), \quad N > N_\nu \right]. \quad (9)$$

For an optical depth of $t = 10$, Fig. 4 shows several examples of the approximation given by Eq. (4) compared to the Monte Carlo (section 4) results and to the plane-parallel approximation [Eq. (3)]. Calculations are shown for two zenith angles, $\theta = 0^\circ$ and $60^\circ$ and for three values of surface reflectance, $a = 0.1, 0.3$ and 0.5.

At $\theta = 0^\circ$, both the Monte Carlo results and the parameterization produce significantly smaller albedo values than does the plane-parallel approximation. At $\theta = 60^\circ$, the Monte Carlo results and parameterization approach the plane-parallel values at high surface albedos.

Of course, real clouds are not cubes, so that the Monte Carlo calculations merely estimate the effects of broken cloudiness. As shown by Welch and Wielicki (1984), the choice of cloud shape leads to differences in cloud reflected fluxes. However, these differences are not expected to greatly modify the shape of the Monte Carlo curves shown in Fig. 4. A more thorough study of cloud field reflected fluxes based upon cloud size distributions is beyond the scope and intent of the present investigation. However, calculations by Aida (1977), Kite (1987), and Cretel et al. (1988) show this to be an important contribution to radiative fluxes which needs to be addressed in more advanced parameterizations.

The parameterization given by Eq. (4) represents a first-order improvement to plane-parallel calculations [Eq. (3)]. The advantages are that the parameterization depends only upon solar zenith angle, cloud optical depth, surface albedo, cloud cover, and plane-parallel cloud albedo. These results are valid only for nonabsorbing clouds. However, surface net radiative fluxes and albedos can be corrected for Rayleigh scattering and water vapor absorption applying the approximations described in Welch and Wielicki (1985). Therefore, spectrally integrated surface radiative fluxes and albedos can be estimated using the previously mentioned techniques making only simple plane-parallel calculations.

6. Discussion and conclusions

In the worldwide effort to monitor climate change, of primary importance are the remotely sensed shortwave and longwave radiances from which the climatically important radiative fluxes are derived. Both local and global climate are governed by a balance between the absorbed solar radiation and the emitted infrared radiation (Ramanathan 1988). Because the effects of global warming are expected to be strongly amplified in polar regions, it is essential to understand better and monitor changes in these regions. However, present remote sensing algorithms are severely compromised by high albedo surfaces, and these surfaces often are incorrectly identified as clouds. In addition, clouds in the Arctic are highly inhomogeneous, requiring broken cloud models in simulations. As pointed out by Maykut and Perovich (1987), the proper treatment of solar radiation absorbed by the water is crucial to the success of modeling ice decay and understanding potential polar climate change.

As shown in Fig. 1, the detection of cloudiness becomes increasingly difficult over higher surface albedos, even at very high spatial resolution. At lower spatial resolution, the brightness of individual cloud contaminated pixels decreases, providing less contrast between clear and cloudy pixels (Minnis and Wielicki 1988). Take, for example, the use of AVHRR GAC data; it is doubtful that broken cumulus cloudiness with about 20% cloud cover is detectable over highly reflecting surfaces of albedo $a \approx 0.25–0.3$. Therefore, the problem is not only one of detecting cloudiness over surfaces covered with ice or snow, but also over deserts and other highly reflecting regions.

The primary result of the present study is the demonstration that not only radiances but also radiative fluxes over high albedo surfaces may be decreased by the presence of broken cloudiness. Conventional wisdom states that cloud radiances (brightness values) are always greater than the background. Most cloud retrieval schemes are built around this fallacious assumption. Figure 1 clearly demonstrates this error. However, that radiative fluxes over a high albedo surface may be reduced by a layer of broken cloudiness is strongly counterintuitive. The resulting increase in surface absorption is shown to be the result of multiple scattering between the surface and the cloud layer. The strength of this multiple scattering contribution is controlled by the probability of radiation reflected by the surface entering a cloud (i.e., cloud field morphology) and by cloud optical thickness.

For a given cloud cover, radiative fluxes are larger at high solar zenith angles than at low solar zenith angles. A fair-weather cumulus cloud field with cloud
cover of 20% composed of moderate optical depth $t = 10$ clouds at $\theta = 0^\circ$ decreases the total solar radiation at the ground by more than 4% over a surface albedo of $a = 0.1$. Over a highly reflecting surface of $a = 0.3$, the decrease is less than 0.5%. This difference is due to the large number of cloud-surface multiple reflections which occur over high background albedos. In such situations, the presence or absence of cloud cover will have little impact upon solar net flux at the surface. However, the impact of broken cloudiness increases with increasing cloud cover, decreasing surface albedo and increasing solar zenith angle (latitude). At $\theta = 60^\circ$, even small cloud covers have an appreciable effect upon reflected solar fluxes and upon the regional radiative energy balance.

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