Satellite estimation of spectral surface UV irradiance
2: Effect of horizontally homogeneous clouds

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Abstract  The local variability of UV irradiance at the Earth's surface is mostly caused by clouds in addition to the seasonal variability. Parametric representations of radiative transfer RT calculations are presented for the convenient solution of the transmission $T$ of ultraviolet radiation through plane parallel clouds over a surface with reflectivity $R_s$. The calculations are intended for use with the Total Ozone Mapping Spectrometer (TOMS) measured radiances to obtain the calculated Lambert equivalent scene reflectivity $R$ for scenes with and without clouds. The purpose is to extend the theoretical analysis of the estimation of UV irradiance from satellite data for a cloudy atmosphere. Results are presented for a range of cloud optical depths and solar zenith angles for the cases of clouds over a low reflectivity surface $R_s<0.1$, over a snow or ice surface $R_s>0.3$, and for transmission through a non-conservative scattering cloud with single scattering albedo $\omega_s = 0.999$. The key finding for conservative scattering is that the cloud-transmission function $C_T$, the ratio of cloudy- to clear-sky transmission, is roughly $C_T = 1 - R_c$ with an error of less than 20% for nearly overhead sun and snow-free surfaces. For TOMS estimates of UV irradiance in the presence of both snow and clouds, independent information about snow albedo is needed for conservative cloud scattering. For non-conservative scattering with $R_s>0.5$ (snow) the satellite measured scene reflectance cannot be used to estimate surface irradiance. The cloud transmission function has been applied to the calculation of UV irradiance at the Earth's surface and compared with ground-based measurements [Herman et al., 1998].

1. Introduction

Changes in cloudiness produce the strongest source of the local variability in the surface UV irradiance [Madronich 1993]. Since any current satellite view is an average over an extended area, the satellite method of mapping surface (UV or PAR) irradiance has to address cloud variability within its field of view FOV. Frederick and Lubin [1988] suggested the use of average UV reflectance within the FOV (170km X 170km) of the Solar Backscatter Ultraviolet (SBUV) instrument to derive UV transmittance through clouds.

The availability of the daily global imagery from the Advanced Very High Resolution Radiometer (AVHRR) instruments aboard the NOAA polar orbiting satellites makes it possible to estimate the cloud transmittance with a higher spatial resolution (~ 2km for local area coverage data and 4km for the global area coverage) than for TOMS observations. The successful applications of the AVHRR techniques to UV-irradiance mapping have been demonstrated so far only for local geographical areas: Germany [Meerkoetter et al. 1997], Antarctic Peninsula [Lubin et al., 1994], and Moscow region, Russia [Rublev et al., 1997]. Although global (GAC) AVHRR data are available (at least for last several years), the technical difficulties (different satellites, calibration issues, large volumes of data, data gaps) prevent using AVHRR data for surface UV irradiance mapping on the global scale.

Another suggested approach is to combine the TOMS ozone data with shortwave cloud reflectance measurements from the NASA Earth Radiation Budget Experiment (ERBE) [Lubin and Jensen 1995; Lubin et al., 1998]. The ERBE data set is available only since 1985 [Lubin 1998]. Also, the transmittance of the cloudy atmosphere is spectrally dependent in the near UV and visible spectral regions [Nack and Green, 1974; Seckmeyer et al., 1997; Krotkov et al., 1997; 2]
Therefore, the extrapolation of the spectrally integrated shortwave ERBE measurements into the UV spectral region is not straightforward and requires further justification.

Surface UV irradiance data derived from TOMS ozone and UV-reflectivity data are based on a single well-calibrated instrument [Eck et al., 1987; Eck et al. 1995]. Using total ozone and 380nm reflectivity measurements from the TOMS instrument aboard the Nimbus-7 satellite, Herman et al. [1996] have estimated the zonal average trends in surface UV flux between January 1979 and December 1992, while neglecting absorbing aerosols. Because of long time record (since 1978) and contiguous global spatial coverage, TOMS data are vital not only for estimating zonal average trends in surface UV flux [Madronich 1992; Herman et al. 1996], but also in studying geographical differences in surface UV climatology, including effects of clouds and aerosols [Herman et al., 1998].

Krotkov et al [1998] examined a UV-mapping algorithm that included aerosol attenuation for cloud- and snow- free conditions, and compared calculated UV irradiances with ground-based data. In the absence of clouds and absorbing aerosols (or for known aerosol amounts), it was demonstrated that satellite measurements of extraterrestrial solar flux, total ozone and UV reflectivity (at 0.36 or 0.38μm) can be used in conjunction with a radiative transfer model to provide estimates of surface spectral UV irradiance to accuracies comparable to that of well maintained ground-based instruments (e.g., Toronto Brewer).

The purpose of the present paper is to extend the theoretical basis of the clear-sky analysis for estimation of surface-UV irradiance from TOMS measured radiances to the case for a cloudy aerosol-free atmosphere with ozone absorption. We also assume a spectrally independent (grey) cloud optical thickness for simulation of the backscattered radiance at the top of the atmosphere and the UV irradiance at the Earth's surface. An issue of concern is the use of satellite measured radiances to estimate an approximation to the hemispherical cloud albedo in the 100x100 km² field of view (FOV). The approximation is based on the assumption that the mixture of cloud-surfaces in the FOV can be represented by a Lambertian equivalent reflectivity (LER) or a plane parallel cloud. The case of a mixture of clear and cloudy scenes within the satellite FOV will be discussed in a future paper.

In section 2 we discuss the relationship between hemispherical albedo of the cloud layer and spectral transmittance of a cloudy atmosphere over snow-free surfaces (low reflectivity). Section 3 uses these results to discuss satellite estimation of the cloud albedo for snow-free conditions. The sensitivity to the presence of snow is discussed in section 4. Section 5 discusses the combining effects of clouds and absorbing (soot) aerosols on the UV atmospheric transmittance.

2. Relationship between cloud albedo and cloud factor

The effect of clouds on global (direct plus diffuse) UV irradiance at the Earth's surface can be described by a relative atmospheric transmittance function, which will be referred to as
the measured Cloud Transmission function \( C_{TG} \) from ground-based measurements (see Equation 1).

\[
F_{\text{cloud}} (\theta_o, O_3, \lambda) = F_{\text{clear}} (\theta_o, O_3, \lambda) \cdot C_{TG}(\theta_o, O_3, \lambda)
\]

\( F_{\text{cloud}} \) is the measured spectral surface irradiance for cloudy conditions, \( F_{\text{clear}} \) is the estimated clear sky irradiance (the same atmosphere but with the clouds removed), \( \lambda \) is the wavelength, \( \theta_o \) is the solar zenith angle.

The calculation of \( F_{\text{clear}} \) from the satellite measurements of the extraterrestrial solar irradiance, total column ozone \( O_3 \) and surface reflectivity \( R_s \) were discussed in Krotkov et al. [1998]. Here we focus on the theoretical estimation of the \( C_T \) using the TOMS satellite reflectivity-channel wavelengths, 360 or 380 nm. For this analysis we define \( C_T \) as the ratio of the calculated atmospheric transmittances using aerosol-free atmospheric models:

\[
C_T(\lambda, \theta_o, O_3) = \frac{T_{\text{cloud}}}{T_{\text{clear}}}
\]

### 2.1 Cloud Without an Atmosphere

Some of the characteristics of the cloud transmission function can be investigated for the simple case of a cloud over a surface with reflectivity \( R_s \). Assuming a horizontally-homogeneous plane-parallel conservative-scattering cloud, and neglecting atmospheric scattering and absorption (\( T_{\text{clear}} = 1 \)), \( C_T \) can be formally expressed as a solution of the Stoke’s problem [Chandrasekhar, 1960]:

\[
C_T = \frac{1 - R_c}{1 - R_s R_{\text{diff}}^C}
\]

where \( R_c \) is the hemispherical albedo of the cloud layer illuminated by direct solar beam for zero surface reflectivity \( (R_s=0) \) and \( R_{\text{diff}}^C \) is the albedo of the same cloud illuminated by diffuse source (see Figure 1). In the near-UV spectral region \( R_s \) is low (2 to 8%) for snow- and ice-free surfaces [Eck et al. 1987; McKenzie et al., 1996; Herman and Celarier 1997]. Neglecting the surface reflection (the general case of arbitrary \( R_s \) will be discussed further in section 4):

\[
C_T(\theta_o, \tau_C, \chi) = [1 - R_c]
\]

For conservative scattering \( R_c \) is a unique solution of the multiple scattering radiative transfer problem for a given cloud phase function, \( \gamma_C \), optical thickness, \( \tau_C \), and solar zenith angle, \( \theta \). The phase function \( \gamma_C \) is determined by the cloud-droplet size distribution. For this analysis we use the CI-cloud model [Deirmendjian 1969]. Figure 1 shows the dependence of \( R_c \) on \( \tau_C \) and \( \mu_0=\cos \theta_0 \) calculated using both numeric (DISORT [Stamnes et al., 1988]) and approximate (two-stream) [Coakley and Chylek, 1975] solutions:
\[ R_C = \frac{\tau_c^*}{(\tau_c^* + 2\mu_0)} \tag{5} \]

where \( \tau_c^* = (1-g)\tau_c \) is the cloud effective optical thickness, \( g \) is the asymmetry factor (mean cosine of \( \gamma_c \); \( g = 0.848 \) for C1 cloud), and \( \mu_0 = \cos \theta_0 \). The accuracy of different two-stream approximations is discussed in many papers [King 1985; Lenoble 1993; Liou 1993]. Here we use Equation 5 and the radiative transfer calculation (DISORT) to illustrate some general properties of \( R_C \): As shown in Figure 1,

- For a given \( \mu_0 \), \( R_C \) is a non-linear monotonically increasing function of \( \tau_c^* \)
- For a fixed \( \tau_c^* \), \( R_C \) is minimal for overhead sun and increases with increase in \( \theta_0 \).

This implies that a cloud layer can reflect a greater percentage of the incident radiation when illuminated by diffuse source than it does for nearly normally incident solar irradiance. The ratio of the upward irradiance to the downward irradiance for diffuse illumination is:

\[ R_C^{\text{diff}} = 2 \int_0^1 R_C(\mu) \mu \, d\mu \tag{6a} \]

\[ R_C^{\text{diff}} = \tau_c^* \left[ 1 - 0.5 \tau_c^* \ln \left( 1 + \frac{2}{\tau_c^*} \right) \right] \tag{6b} \]

Figure 1 Cloud albedos \( R_c \) and \( R_c^{\text{diff}} \) as a function of \( \tau_c \) or \( \tau_c^* \) and \( \mu_0 = \cos(\theta_0) \). The discrete points are from the DISORT calculation and the continuous lines are from the Equations 5 and 6.
One important consequence of $R_{c}^{\text{Dir}} > R_{c}(\mu_{0}=1)$ is the theoretical possibility of apparent enhanced cloud transmittance ($C_{T}>1$) over bright snow surface (discussed later in section 4), since the top of the cloud is illuminated by the direct sun while the bottom of the cloud is illuminated by diffuse radiation reflected from the snow [Shettle and Weinman, 1970].

If $\tau_{c}$ is parameterized in terms of the column liquid water content, $W$ (usually in units of grams per m$^{2}$), and the effective radius of the droplet size distribution, $r_{\text{eff}}$, then $\tau_{c} \approx 1.5W/r_{\text{eff}}$ [Stephens, 1978]. For a given water content and solar zenith angle ($W=\text{const}$, $\mu_{0}=\text{const}$), increasing the effective radius of cloud droplets results in increasing cloud transmission $C_{T}$ (see Figure 1 and Equation 5).

Even though we neglect spectral dependence of $\tau_{c}^{*}$, the strong spectral dependence of Rayleigh scattering ($\sim \lambda^{-4}$) and ozone absorption in the UV-B spectral region (290 to 320 nm) causes the combined relative transmittance $C_{T}$ of the cloud plus atmosphere (Equation 2) to be spectrally dependent and usually larger than 1- $R_{c}$ (from Equation 4). In the next section we consider these atmospheric effects.

### 2.1 Atmospheric Effects

To study the sensitivity of $C_{T}$ to atmospheric parameters and the relationship with cloud co-albedo (1-$R_{c}$), we consider a model that includes a homogeneous water-cloud layer of optical depth $\tau_{c}$ embedded in a scattering molecular atmosphere with ozone absorption.

Using calculated values of $T_{\text{clear}}$ and $T_{\text{cloud}}$, $C_{T}$ is calculated from Equation (2) in the UV spectral region (290-400nm) for a wide range of input atmospheric parameters and solar zenith angles $\theta_{0}=0-70^{0}$. For a cloud- and aerosol-free atmosphere with a Lambertian reflecting surface, the atmospheric transmittance $T_{\text{clear}}$ is calculated for each wavelength from

$$T_{\text{clear}} = (T_{\text{dir}} + T_{\text{diff}}) / (1-R_{S}S_{b}), \quad (7)$$

where $T_{\text{dir}}$ and $T_{\text{diff}}$ are the direct and diffuse irradiance at the ground for unit solar flux and zero surface reflectivity [Chandrasekhar, 1960]. The factor $(1-R_{S}S_{b})^{-1}$ accounts for the effect of surface reflectivity, where $S_{b}$ is the fraction of reflected radiation backscattered to the surface by the cloud-free atmosphere and $R_{S}$ is the surface reflectivity (albedo). We calculate the various terms in Equation 7 for a Rayleigh atmosphere with standard ozone profiles [McPeters et al., 1996; Wellemeyer et al., 1997] using numerical solutions of the DISORT radiative transfer equation for large optical thicknesses where polarization can be neglected, and for modified versions of the Gauss-Seidel code [Herman and Browning 1965; Ahmad and Frazer, 1982] for smaller optical depths where polarization can have a noticeable effect on radiances at the ground and the top of the atmosphere.

The high-spectral resolution (0.05nm) temperature-dependent ozone-absorption coefficients are based on the laboratory measurements of Bass and Paur [1985] and the Rayleigh scattering coefficients from Bates [1984]. We use the same atmospheric model with an
embedded horizontally homogeneous water cloud to calculate $T_{\text{cloud}}$. The cloud parameters are the cloud base and top altitudes, cloud optical thickness $\tau_c$, and the cloud phase function $\gamma_c$. $\tau_c$ is assumed spectrally independent and $\gamma$ corresponds to C1-cloud model [Deirmendjian 1969]. The effective phase function is determined as a mixture of the cloud and Rayleigh phase functions for each computational layer $\Delta\tau_c$, and is wavelength dependent through the scattering mixing ratio ($\Delta\tau_c / \Delta\tau_{\text{Rayleigh}}(\lambda)$).

2.2.1 $C_T$ Spectral Dependence

Figure 2 shows the calculated spectral dependence of atmospheric transmittance for clear ($T_{\text{clear}}$) and cloudy ($T_{\text{cloud}}$) conditions as well as $C_T(\lambda)=T_{\text{cloud}}/T_{\text{clear}}$.

![Diagram showing spectral dependence of atmospheric transmittance](image)

**Figure 2** Spectral dependence of $C_T$, $T_{\text{clear}}$, and $T_{\text{cloud}}$ (290nm-400nm) for $\tau_c=10$.

Both $T_{\text{cloud}}$ and $T_{\text{clear}}$ decrease at shorter UV-A wavelengths (320-400nm) due to increasing molecular scattering ($\tau_R(\lambda) \sim \lambda^{-1}$). Ozone absorption is weak in the UVA spectral region, but increases exponentially at shorter UVB wavelengths ($\lambda<320$nm). Because of stratospheric ozone absorption (maximum ozone concentration at 25 km) the atmospheric transmittance drops several orders of magnitude at shorter UVB wavelengths. The effect of stratospheric ozone absorption on $C_T$ is canceled by calculating the ratio $T_{\text{cloud}} / T_{\text{clear}}$ for the same ozone amount. As a result, $C_T(\lambda)=T_{\text{cloud}} / T_{\text{clear}}$ is a weak function of wavelength in the UVA.
(320nm-400nm) spectral region, and only shows ozone effects from multiple scattering within the cloud for the shortest wavelengths (maximum ozone absorption). Using $C_T$ to estimate cloud transmittance in the UV is useful because of the independence of $C_T$ from stratospheric ozone absorption, with only a weak dependence on tropospheric absorption at the shortest wavelengths of interest ($\lambda < 300$ nm), except for clouds with greater altitude thickness (e.g., thunderstorms, Mayer et al., 1998)

The residual spectral dependence of $C_T$ (~20%) (see Figure 2) is due to second order effects, which remain after the cancellation of the first order effects (i.e. stratospheric ozone absorption). Multiple reflections between the cloud layer and Rayleigh atmosphere tend to increase $C_T$ much like reflection from the ground (Equation 3). On the other hand, enhancement in tropospheric ozone absorption in the presence of clouds tends to decrease $C_T$. As a result, the spectral dependence of $C_T(\lambda)$ has a broad maximum at 300-330nm, with a decrease longward of the maximum (due to decreasing of $s_r(\lambda) \lambda^{-4}$), and a sharp decrease shortward (due to the spectral enhancement of absorption by tropospheric ozone).

To illustrate the behavior of $C_T(\lambda)$ it is convenient to parameterize the spectral dependence by normalizing at a specific wavelength $\lambda_0$:

$$C_T(\lambda) = S_c(\lambda, \theta_o, \tau_c) C_T(\lambda_0)$$

(8)

Where $S_c(\lambda) = C_T(\lambda)/C_T(\lambda_0)$, and $\lambda_0$ is any UVA wavelength, where ozone absorption is negligible. Figure 3 shows the wavelength dependence of $S_c(\lambda, \theta_o, \tau_c)$ for $\theta_o = 0^\circ$ and $70^\circ$ and $\tau_c = 1, 3, 5, \text{ and } 10$. This shows that the wavelength dependence increases strongly with increasing $\theta_o$ and with increasing $\tau_c$. Thus, the atmosphere/cloud system, with a spectrally independent $\tau_c$ becomes relatively more transparent in the UVB than in the UVA (320-400nm) and visible spectral regions [Nack and Green, 1974; Seckmeyer et al., 1997; Krotkov et al., 1997; Frederick and Erlick, 1997].
2.2.2 $C_T$ Dependence on Cloud Optical Depth

For overhead sun and a dark surface below the cloud, Figure 4 shows that the dependence of $C_T$ on $\tau_c$ is roughly the same as the dependence of $(1-R_C)$ on $\tau_c$ (see Equation 5). This is consistent with the assumed cloud transmittance used in previous studies [Eck et al., 1995; Herman et al., 1996]. At weakly ozone absorbing wavelengths (315-400nm) $C_T$ is larger than $(1-R_C)$ due to the Rayleigh backscatter effect from the atmosphere above the cloud. The increase is 10 to 12% at $\tau_c=20$. However, at strongly ozone absorbing wavelengths (<300nm), $C_T$ becomes smaller than $(1-R_C)$, due to enhancement in tropospheric ozone absorption compared to longer wavelengths. At 294 nm and $\tau_c=20$ the decrease is about 13%. In the limiting cases of very small ($\tau_c<1$) or extremely high ($\tau_c>100$) cloud optical depths, $C_T$ approaches $(1-R_C)$ for wavelengths where there is little or no ozone absorption. When there is ozone absorption (see Figure 4 for 294 nm) and the cloud optical depth $\tau_c>1$, tropospheric ozone absorption is enhanced within the cloud leading to deviations from 1-$R_C$. The optical depth at which $C_T$ deviates from 1-$R_C$ is a function of wavelength. The geometric extent of the cloud is also important, especially at short UV wavelengths [Meyer et al., 1998].

![Figure 3](image-url)
2.2.3 $C_T$ Dependence on Solar Zenith Angle

In the limiting case of visible wavelengths, one can neglect the effects of atmospheric scattering and ozone absorption, thus $C_T \sim (1-R_c) / (\tau + 2 \mu_o)$ (i.e., $C_T$ increases with increasing $\mu_o$, i.e., the maximum $C_T$ is for overhead sun). For wavelengths where atmospheric-multiple scattering and absorption must be considered (UV-A and UV-B), multiple reflections between the cloud and molecular atmosphere effectively inhibit this dependence, so that $C_T$ is a weak function of solar zenith angle. For $C_T(\lambda > 300 \text{nm})$, $dC_T/d\theta_o < 0$ or approximately $-6.7 \times 10^{-4}$ for $\lambda = 310 \text{nm}$, and for strongly ozone absorbing wavelengths ($\lambda < 300 \text{nm}$) the $C_T$ dependence on solar angle is reversed, $dC_T/d\theta_o > 0$ or approximately $5 \times 10^{-4}$ per degree for $\lambda = 294 \text{nm}$.

The sensitivity study conducted in this section has shown that for low surface albedo the cloud optical depth is the single most important parameter determining the value of $C_T$ and the surface irradiance for overcast conditions. Since the $C_T$ dependence on $\tau_c$ is approximately the same as the dependence of $(1-R_c)$ on $\tau_c$ (see Figure 4) for small solar zenith angles, $C_T(\theta_o < 20^\circ)$ can be approximated directly from the satellite estimation of cloud albedo without inferring cloud optical depth as an intermediate step. Second order effects ($C_T$ spectral dependence and dependence on solar zenith angle) are well understood and can be captured from the
homogeneous cloud model. Therefore, for satellite estimation of UV transmittance through clouds, the problem of estimating $C_T$ is reduced to the estimation of cloud hemispherical albedo at non-ozone absorbing wavelengths. In section 3 we will discuss this problem in detail.

![Figure 5](image)

**Figure 5** The dependence of $C_T$ on $\theta_o$.

3. **Satellite Estimation of Cloud Factor $C_T$**

In a satellite algorithm designed to estimate cloud albedo, $R_c$ must be derived from measured backscattered radiances at the top of the atmosphere (TOA). To achieve this goal, three main problems should be considered:

- Subtraction of atmospheric (Rayleigh) backscatter above the cloud;
- Model of the angular (non-Lambertian) distribution of the radiation at TOA;
- Subtraction of the surface reflectance from the total scene reflectivity.

Generally, the first and last problems are coupled. Fourteen years of TOMS radiance data and RT modeling have been used to develop a climatology of snow-free surface reflectivity [Herman and Celarier, 1997]. For low UV surface albedo typical of snow-or ice-free surfaces, or an optically thick cloud layer, one can neglect this coupling as a first approximation. In this section we consider two RT models developed to estimate cloud albedo over low-reflection surfaces ($R_s < 0.1$) from the satellite single-channel radiance measurements. The general case of highly reflecting surfaces will be considering in section 4. We start with discussion of the Lambert Equivalent Reflectivity (LER) model [Eck et al., 1995] (Section 3.1). Next we discuss a single layer horizontally homogeneous cloud model (Section 3.2). Until Section 5 we consider strictly conservative cloud scattering (single scattering albedo $\omega_o = 1$).
3.1 Lambert Equivalent Reflectivity Model

The backscattered radiance emerging from the top of atmosphere as seen by the satellite instrument, $I_T$, can be expressed as a sum of the atmospheric backscatter above the cloud, $I_o$ and the radiance reflected from the top of the cloud [Chandrasekhar, 1960; Dave, 1978; Eck et al., 1987; Eck et al., 1995; Herman et al., 1996]:

$$I_T(\theta, \theta_o, \phi_c, R_T) = I_o + R_T T(\theta, \theta_o, \phi_c) / [1 - R_T S_b(\phi_c)]$$ (9)

where $R_T$ is the Lambert Equivalent Reflectivity LER, $S_b$ is the diffuse reflection of Rayleigh atmosphere illuminated from below by an isotropic source, $\theta$ is the satellite scan angle, $T$ is the transmission function of a Rayleigh atmosphere, and $\phi_c$ is the cloud-top pressure. For the wavelengths of interest, $I_o$, $S_b$, and $T$ are calculated assuming a purely Rayleigh atmosphere with no gaseous absorption and no Mie scattering. (A small correction is made for Rotational-Raman scattering, using the procedure described by Joiner et al. [1995]). $\phi_c$ can be estimated from the monthly satellite cloud climatology data set as a function of location and season (International Satellite Cloud Climatology Project (ISCCP), Rossow and Schiffer, 1991).

Using $R_T$ to estimate $T_{CLOUD}$ assumes that the effect of clouds on the backscattered radiances in UV is similar to that of a Lambertian reflecting surface with enhanced albedo [Eck et al., 1987; Eck et al., 1995; Herman et al., 1996; Krotkov, 1997]. It should be stressed that the $R_T$ is calculated for each scanning direction to fit the satellite measured radiance. If the cloud reflects radiation isotropically, the $R_T$ would be the true cloud hemispherical albedo, $R_T = R_c$. In general, clouds are non-Lambertian reflectors (e.g., for empirical ERBE angular dependence models see Suttles et al. [1989]).

Because of the angular-dependent reflectance of the clouds systems, $R_T$ is a function of the satellite zenith angle, $\theta$, solar zenith angle, $\theta_o$, and the azimuthal angle, $\phi$. For a clear scene, $R_T$ is close to the ground plus tropospheric aerosol reflectivity, which is uniformly low $R_s < 0.1$ in the UV spectral region [Eck et al., 1987; Herman and Celarier, 1997].

In an earlier work [Eck et al., 1995], it was assumed that the UV reflectivity, $R_T$, measured by the Total Ozone Mapping Spectrometer (TOMS) is representative of the cloud hemispherical albedo with a small correction for the underlying surface reflectivity so that $R_T - R_s$ could be used to estimate the $C_T$.

Figure 6 shows the relationship between $R_T$ at 380nm and the cloud hemispherical albedo, $R_c$ (see Equation 1) for a 1 km thick cloud embedded in a mid-latitude ozone containing atmosphere, zero surface albedo, and typical Nimbus-7/TOMS scanning geometry (Figure 6a shows nadir- and Figure 6b the 54° (azimuth 90°) viewing directions). The results show that although there is a general correlation between $R_T = R_{380}$ and $R_c$, these quantities are not the same. The differences between $R_c$ and $R_T$ depend on the viewing geometry, particularly the solar zenith angle.
Figure 6  \( R_{380} \) vs \( R_c \), \( \theta_o=0^\circ \) (solid line) and \( \theta_o=60^\circ \) (dotted line). a) Left Panel: nadir observational direction \( \theta=0^\circ \); b) Right Panel: \( \theta=54^\circ \); \( \varphi=90^\circ \).

Estimates of \( C_T \) with this method are most accurate for the Nimbus-7 TOMS mid-latitude scanning geometry (TOMS overpass near local noon and the instrument scan nearly perpendicular to the principal plane of the sun), which minimizes the effects of cloud anisotropic angular reflectance (Figure 6b). However, for other conditions (high latitudes), the failure to account for angular dependence of the backscattered radiation field can lead to large errors in this method. For common observing conditions the errors can be 20\% or more.

Neglecting the surface reflectivity \( R_s \) is justified in the case of the optically thick clouds (\( R_c > 0.5 \)). For optically thin clouds over a low-reflectance surface, a simple correction for the underlying surface contribution was proposed [Eck et al., 1995]:

\[
R_{ECK} = \frac{(R_{380} - R_s)}{(1-2R_s)}
\]

\[
C_{ECK} = 1 - R_{ECK}, \quad R_{380} < 0.5
\]

\[
C_{ECK} = 1 - R_{380}, \quad R_{380} > 0.5
\]

where \( R_s \) is based on minimum \( R \) values measured by Nimbus-7/TOMS [Eck et al., 1995; Herman and Celarier, 1997].

The simple \( R_{380} \)-LER method was compared with measurements at Toronto, Canada for mid-latitude observational conditions using direct comparisons with the ground based Brewer #14 UV irradiance measurements [Eck et al., 1995]. Reanalysis of this data [Herman et al., 1998] showed that there is a 20\% disagreement (TOMS estimates greater than Brewer measurements) even though estimates of UV irradiance are made near local noon with the TOMS instrument scanning nearly perpendicular to the principal plane of the sun. The 20\%
disagreement includes $C_T$ errors in addition to other systematic instrument or aerosol related errors.

3.2 Plane-parallel cloud model

To implement an improved cloud correction compared to Equation 10, detailed radiative transfer calculations were done to calculate the downward irradiance at the ground, and the upward radiances at the top of the atmosphere, using a midlatitude 325 DU ozone profile and a homogeneous Diermendjian [1969] "C-1" type cloud located between 3.5 km and 5.0 km. The calculations were conducted with cloud-optical thicknesses from 0 to 100, surface reflectivities from 0 to 1, surface heights of 0 and 3.5 km, and solar zenith angles from 0 to 88°. Upward radiances were post-processed into effective cloud optical depth [Krtokov et al., 1997; Herman et al., 1998].

We should stress, however, that TOMS UV technique does not rely on retrieval of the real cloud optical thickness (i.e., total column water content). The effective $\tau_{\text{eff}}$ derived from TOMS radiances would be equal to the real $\tau_c$ only in the idealized case of a plane-parallel horizontally-homogeneous cloud layer. In the real case, $\tau_{\text{eff}}$ becomes a function of the assumed surface reflectivity, TOMS observational geometry, cloud fraction, and solar zenith angle. Neglecting cloud fraction, a lookup table for $C_T$ was built that gives $\tau_{\text{eff}}$ ($\theta_s, R_T, R_s$). The value of $R_s$ was taken from a climatological database that was developed using the 15 years of TOMS data [Herman and Celarier, 1997]. As a final step, a Lagrange interpolation was used to determine the $C_T$ between the tabulated parameter values using the $\tau_{\text{eff}}$ as a parameter.

![Figure 7](image)

Figure 7. Cloud transmittance factor, $C_T$ at 325nm as a function of the 380nm Lambert Equivalent Reflectivity R380, for solar zenith angles between 0° and 70°, derived from TOMS backscattered radiance.
Figure 7 shows $C_T$ at 325 nm as a function of $(1 - R_{380})$ for a 1-km thick cloud embedded into a mid-latitude ozone containing atmosphere at 5 km altitude. Compared to $(1 - R_{380})$, the values of $C_T$ are about 20% larger for thick clouds ($\tau_c > 10$) and $\theta_0$ less than $-40^\circ$. Similar calculations at different measurement geometries (see Herman et al., 1998) indicate that equation (10) can be used to estimate $C_T$ to about 10% - 20% accuracy for snow-free conditions. Since the LER model does not give the spectral dependence of $C_T$ (see Figure 2), the LER method would underestimate $C_T$ at biologically important UV-B wavelengths (e.g., 310nm).

3.3 Comparisons with Ground-Based Data Sets

Satellite data represent instantaneous radiance measurements averaged over the large geographical extent in the footprint (the TOMS footprint is on average 100 km by 100 km). Because of high cloud variability and different FOV between ground and satellite instruments, only time-averaged irradiances should be compared to ground-based measurements. The period of averaging should be at least a week in order to minimize the effects of local cloud changes. Herman et al. [1998] compared the 7 days running averages of the daily TOMS estimated UV-irradiances with corresponding 7-day running averages of the daily Brewer (#14) irradiances in Toronto (obtained within ±30 min of TOMS overpass time). They found that during the summertime the TOMS estimated irradiances are larger by about 20% than those measured by Brewer spectrometer. These differences represent the combined effect from different cloud and aerosol conditions including calibration uncertainties of both instruments [Herman et al., 1998].

An alternative approach is to group the measurements according to satellite reflectivity data and compare the average irradiances for each reflectivity range. Comparing only cloud factor data $C_T$ (after normalizing the irradiance by the clear sky irradiance) reduces the instrument-calibration bias (mainly the cosine correction [Krotkov et al., 1998]), to the extent that it is the same for clear and cloudy conditions. This method is useful for understanding differences between the satellite algorithm and ground-based measurements for varying cloud amounts and corresponding average reflectivity in the TOMS FOV.

The daily TOMS estimates of $C_T$ (sections 3.1 and 3.2) were compared with ground-based Brewer spectrometer measurements of the same quantity. The Brewer spectra were measured at Toronto between 1989 and 1998 by instrument #14. The comparisons were done under conditions without snow or ice (May through September data). The Brewer cloud factor ($C_{TB}$) was obtained by dividing the measured irradiance at 324 nm by "statistical clear sky" Brewer-irradiance parameterization [Fioletov 1997]:

$$T_{\text{CLEAR}(324)} = -0.119848 - 0.0239033 \theta - 0.00124988 \theta^2 + 0.000274423 \theta^3$$
$$- 1.5479E-005 \theta^4 + 4.20862E-007 \theta^5 - 6.10836E-009 \theta^6$$
$$+ 4.54795E-011 \theta^7 - 1.36825E-013 \theta^8$$ (11)
where θ is the solar zenith angle in degrees. The parameterization is based on a 95-percentile method: when the measured irradiance data (after the Earth-Sun distance adjustment) are compared to this formula, for each solar zenith angle approximately 95% of all data are less than \( T_{\text{CLEAR}}(324) \) and 5% are more than \( T_{\text{CLEAR}}(324) \).

Figure 8 Brewer #14 measured \( C_{TB} \) versus TOMS measured \( R_{380} \) from daily measurements in Toronto during May to September, 1989-1993. a) The left panel shows the results of reflectivity binning, and b) the right panel shows a 20-point running average to the data sorted by reflectivity \( C_{20}^{TB} \) and a the linear least squares fit to the data points \( C_{TB}^{\text{FIT}} \). The dashed line is \( X=C_{ECK}=1-(R_{380}-0.05)/0.9 \) [Eck et al., 1995].

For Toronto during May to September θ is between 20° and 50° for the near-noon TOMS overpass times. The new cloud factor \( C_t \), shown in Figure 7 (TOMS nadir view), is larger than \( 1-R_{380} \) for cloudy skies \( (R_{380} > 0.5) \) and approaches \( 1-R_{380} \) as \( R_{380} \) becomes small (near \( R_s \)). For cloudy skies \( C_t \) is similar to \( C_{TB} \), but \( C_t > C_{TB} \) for nearly clear skies. The clear sky discrepancy has been interpreted in terms of absorbing aerosols that are seen by the Brewer instrument, but
not by TOMS [Krotkov et al., 1998; Herman et al., 1998]. The presence of absorbing aerosols would make $C_{TB} < 1$ when $R_{380}$ becomes small.

The Brewer measured $C_{TB}$ data (measured within $\pm 30$ of TOMS overpass time) were binned using 10% intervals of TOMS $R_{380}$ to calculate means and standard deviations. The mean Brewer $C_{TB}$ values for each $R_{380}$ bin are shown in Figure 8a. The error bars are $\pm 2\sigma$ ($\sigma$ is the standard error of the mean) for each bin. Running averages of $C_{TB}$ and a least squares fit to the data points are shown in Figure 8b (1-hour average) and 8d (4-hour average). The dashed line shows the reflectivity-based TOMS estimated cloud correction factor for surface reflectivity 0.05 [see Equation 10 and Eck et al. 1995]. For both 1-hour and 4-hour cases, TOMS overestimates $C_T$ for broken and thin clouds ($R_{380} < 0.3$), but underestimates for thick clouds ($R_{380} > 0.45$). The running average $C_{TB}$ shows that the slope of the least squares fit is locally meaningful and not determined by the large cluster of points where $R_{380}$ is near 1.

Figures 8b and 8d show that $C_{ECK}$ overestimates $C_T$ for broken and thin clouds ($R_{380} < 0.3$), but underestimates for thick clouds ($R_{380} > 0.45$). At least part of the Eck et al [1995] underestimation for thick clouds is due to neglecting of $C_T$ spectral dependence (see section 2.2.1), which has been recently incorporated into the current TOMS cloud correction algorithm (see figure 5 where the difference between the $C_{ECK}$ and current TOMS $C_T$ [Krotkov et al., 1997; Herman et al 1998] is shown).

Figure 8 shows that $R_{380}$ provides an estimation for $C_{TB}$ with some systematic bias, and that the random error of any single measurement is large (see Figure 8b). The random error results mainly from the fact that one compares the spatially averaged satellite data with a single-point ground measurement under highly variable cloud conditions. However, Figure 8 also shows that the random error can be reduced by considering averages over reflectivity bins or by using running averages (this is similar to the effect obtained using weekly averaging [Herman et al., 1998]). The remaining differences are systematic and represent the bias of particular satellite cloud correction algorithm.

Brewer measurements and TOMS irradiance estimates are based on different cloud amounts in their respective fields of view. For example, when $R_{380}$ is small, the sky is most likely to be filled with a mixture of clear and cloudy regions. At the TOMS overpass time the Brewer instrument can measure an irradiance value that is greater or less than the TOMS value with nearly equal probability. Increasing the time over which the Brewer data are averaged increases the likelihood that the average cloud amount in the TOMS and Brewer scenes are similar. This can be seen by comparing Figures 8b and 8a. Figure 8b contains 1 measurement per point and 8d contains 4 Brewer measurements that are averaged to produce 1 data point. For low $R_{380}$ the spread of $C_{TB}$ about the $1 - R_{380}$ is significantly reduced in the 4 hour average case (Figure 8d). Standard deviations are 37%, 26%, and 21% for 1-hour, 2-hour and 4-hour averaging, respectively. A 6-hour average has larger standard deviations than the 4 hour average because the difference in time between Brewer and TOMS measurements is too large.
Figure 9  A comparison between monthly averages of 4-hour daily averages of the Brewer $C_{TB}$ and the TOMS $1-R_{380}$ [ •️], and a comparison between 5-month running averages $<X>$. The boxed symbols [ •️] are averages for the same month that are close in value.

When the 4-hour daily data are combined into monthly averages (see Figure 9), the agreement between the TOMS and Brewer estimated cloud transmittances is improved (each pair of points within a box is for the same month). The data used are for the period 1989 to June 1998 and are from both Nimbus-7/TOMS (1979 to May 1993) and Earth-Probe/TOMS (August 1996 to present). For most months the agreement is sufficiently close to be useful for calibration of the ground-based Brewer system. Since some months are too much in disagreement for calibration purposes, it is recommended that running averages of the monthly averages be used. Here the agreement is within a few percent, well below the instrument to instrument uncertainty (10 to 20%).

4. Clouds Over Snow

The UV and visible reflectivity of fresh snow can be extremely high: $R_s > 0.95$ [Bohren and Barkstorm 1974; Wiscombe and Warren 1980; Grenfell et al., 1994; Zege and Kokhanovskii, 1997; Herman et al., 1998]. Therefore, the possible errors in satellite-$C_T$ estimation due to surface-albedo uncertainty differ significantly for winter and summer conditions. Here we consider the case of conservative cloud scattering, but allow arbitrary surface albedo, $R_s$. The basic behavior of the cloud-surface system can be understood while neglecting atmospheric scattering. This approximation is useful for $\lambda \geq 360$ nm where the atmospheric scattering effects
are small. Including the atmosphere causes the 1-R_{\text{SYSTEM}} lines to be replaced with curved lines (see Figure 4). In addition, R_{\text{SYSTEM}} can be greater than 1 for some observing angles. This effect is observed in the TOMS data.

For arbitrary R_s and neglecting atmospheric effects, C_T can be expressed as a solution of the Stokes problem (Equation 3). The albedo of the cloud-surface system, R_{\text{SYS}}, can be calculated from the energy balance:

\[ R_{\text{SYS}} = R_C + T_C R_s T^* / [1 - R_s R_{C}^{\text{dir}}] \]  

where \( T_C = 1 - R_C \) and \( T^* = 1 - R_{C}^{\text{dir}} \) are cloud transmittance of direct and diffuse radiation for a black underlying surface. Equations (3) and (12) present a formal relationship between C_T and albedo of the system for arbitrary R_s. (See Figure 2 and Equation 4 for values of R_{C}^{\text{dir}}).

Figure 9 illustrates the relationship between C_T, R_{\text{system}} and R_s for a homogeneous plane-parallel cloud model for \( \theta = 0^\circ \) and \( 60^\circ \) in the absence of a scattering atmosphere. C_T was calculated directly from its definition (Equation 2) for different values of surface albedo using the DISORT radiative transfer program. The diagram confirms our qualitative conclusions and allows quantitative C_T estimation for any combination of R_{\text{SYSTEM}} and R_s. It is quite obvious that in the presence of snow (R_s>0.3) one needs to know both measured reflectivity R_{\text{SYS}} and surface albedo R_s, to estimate C_T (even for thick clouds, \( \tau_C > 30 \)). These two independent quantities cannot be estimated from a single channel satellite reflectivity.
Figure 9  The relationship between the total Cloud Factor $C_T$ and co-albedo of the “cloud-surface” system, $1-R_{\text{system}}$ using DISORT [Stamnes et al., 1988] for a conservative cloud layer [C1, Deirmendjian, 1969] over a Lambertian reflective surface with reflectivity $R_s$ between 0 and 0.95 (fresh snow). a: Top) Overhead sun: $\theta_0=0^\circ$;  b: Bottom) $\theta_0=60^\circ$
For TOMS, the best that can be done for estimating UV transmittance through clouds over snow/ice, in the absence of independent information, is to use a climatology for estimating $R_s$. The fraction of the measured reflectivity associated with a possible cloud is assigned on a probabilistic basis. For example, in a suburban area with trees and buildings, the surface reflectivity over a region of 100x100 km$^2$ (TOMS FOV) rarely exceeds 0.4 when there is snow on the ground (except immediately after a fresh snowfall). Measured reflectivities in excess of a threshold value could be assigned to an overlying cloud layer. For estimation of UV irradiance, averaging the transmittance over at least a week should minimize errors caused by fresh-snow conditions compared to the true weekly dosage.

For fresh snow ($R_s>0.5$) and $\tau<50$ and overhead sun the irradiance at the surface exceeds that for a clear atmosphere (Figure 9a) [see also Shettle and Weinman, 1970]. If $R_s > R_c/R_c^{\text{Def}}$, then $C_T>1$ (see Equation 3 and Section 2.1). Using a two-stream approximation for $R_c$ (see Equation 5), the condition $C_T>1$ can be expressed as:

$$R_s > 2[\left(\tau^*+2\mu_0\right)\ln(1+2/\tau^*)]^{-1}$$

(13)

From this equation one can see that a combination of low solar-zenith angle ($\mu_0\sim1$) and high surface albedo is required for enhanced cloud transmittance ($C_T>1$). In this regime the cloud layer becomes more transparent from above (direct solar radiation) than from below (diffuse illumination by the snow surface). Therefore, photons are "trapped" between the surface and the cloud bottom ("cavity" effect). The cloud optical depth can vary in a wide range (e.g., $\tau_c =1$ to 50). We note that $C_T>1$ regime does not violate energy conservation, which can be written in the form:

$$C_T(1-R_s)= R_{\text{system}}$$

(14)

The condition of low solar-zenith angle ($\mu_0\sim1$) is critical. As can be seen from Figure 9b, the enhanced transmittance ($C_T>1$) does not occur for $\mu_0=0.5$ ($\theta_0=60^\circ$). Also, optically thick clouds eventually reduce $C_T$ to less than unity ($\tau>50$).

5. Non-Conservative Cloud Scattering

In the satellite estimation of cloud transmittance in the visible and near UV spectral regions ($\lambda \geq 360$ nm), a conservative-scattering regime is usually assumed ($\omega=1$). Calculations [Twomey, 1972; Twomey and Bohren, 1980; Twohy et al., 1989; Chylek et al., 1996] and direct measurements [King et al., 1990] of the cloud single-scattering albedo show that this is a reasonable assumption for clean marine clouds ($\omega>0.9999$). Yet other measurements (although not direct) estimate $\omega < 0.999$ at 0.4$\mu$m [Melnikova et al., 1997]. In this section we extend our analysis to the case of slightly non-conservative cloud scattering (which might be the case for a mixture of clean-water droplets with absorbing aerosols such as those containing soot). Since we are considering wavelengths where scattering effects are not important, we neglect the atmosphere in the model calculations.
Figure 10 is similar to Figure 9 except the cloud scattering is assumed slightly non-conservative ($\omega=0.999$). We note that differences are not large for small $\tau < 10$ and low surface reflectivity ($R_s<0.1$). However, the general relationship between $C_T$ and $R_{\text{system}}$ is quite different from conservative case (Figure 9a). For example, for low $R_s<0.1$, $R_{\text{system}}$ increases with $\tau_c$, but never gets larger than $\sim 80\%$ even for $\tau_c>100$. In other words, even very thick $\omega=0.999$ clouds do not appear as bright as thinner $\omega=1$ clouds. On the other hand, over fresh snow surfaces, the reflectivity of the scene decreases with increase in cloud optical thickness because of multiple reflections and losses within the cloud. For $R_s<0.8$, $R_{\text{system}}$ does not change with $\tau_c$, but $C_T$ does ($R_{\text{system}}=0.8$ while $C_T$ changes from 1.1 to zero with increase in $\tau$). Thus, the satellite reflectance measurements are not useful for determining $C_T$ for $R_s>0.5$ even if $R_s$ is known a-priori. Enhanced cloud transmittance regime ($C_T>1$) is still possible even for slightly non-conservative cloud scattering. We have shown that even slight absorption makes it impossible to estimate $C_T$ over bright snow surfaces from satellite reflectance measurements.

![Figure 10](image)

Figure 10  Similar to Figure 9a but for non-conservative scattering, $\omega_o=0.999$

Although the details are not shown in this paper, the spectral dependence of a slightly absorbing cloud is quite different than for the case where $\omega_o=1$. 

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6. Summary

Radiative transfer calculations have been presented for the convenient estimation of the transmission \( T \) of ultraviolet radiation through plane parallel clouds over a surface with reflectivity \( R_s \). For snow-free surfaces and conservative-cloud scattering it was demonstrated that the transmittance of a cloudy atmosphere can be approximated from the satellite measurements of reflectivity \( R_c \) at non-absorbing UV wavelengths as \( 1 - (R_c - R_s) \) with an error of about 20\%. The amount of error depends most strongly on the solar zenith angle and optical depth. At this level of approximation, the TOMS \( R_{SYS} \) measurements can be utilized directly to approximate cloud albedo, cloud transmittance or \( C_T \), and the irradiance at the Earth’s surface.

A better approximation to \( C_T \) can be obtained from an atmospheric radiative transfer RT model with homogeneous-cloud layer to convert satellite measured radiance to cloud transmittance using a table look-up method. The RT method predicts larger cloud transmittances than the \( 1 - (R_c - R_s) \) method for solar zenith angles \( \theta_s < 50^\circ \), and accounts for the spectral dependence of the cloud factor \( C_T \). Radiative transfer calculations show that the effects of cloud anisotropic reflectance from non-Lambertian surfaces are reduced in the plane perpendicular to the solar-principal plane. This is the geometry appropriate for the near-noon radiance measurements obtained from TOMS.

The cloud transmittance function \( C_{TB} \) has been derived from Brewer #14 data in Toronto, Canada for May to September 1989 - 1993 and compared to the satellite estimations of \( C_T \). As expected, the day to day variations in reflectivity are too large to be directly compared to \( C_T \). When the Brewer data are averaged in time or in reflectivity bins it has been shown that \( C_{TB} \) is close to \( 1-R_c \) and that there are 20% systematic differences relative to \( 1-R_c \) for clear and cloudy conditions.

For TOMS estimates of UV irradiance in the presence of both snow and clouds, independent information about snow albedo is needed for conservative cloud scattering. However, for nonconservative cloud scattering (e.g., \( \omega = 0.999 \)) with snow reflectance above 0.5, the satellite measured scene reflectance cannot be used to estimate surface irradiance. Adding atmospheric scattering to the model does not change this conclusion. For low solar-zenith angles and high surface albedo, enhanced cloud transmittance is possible (\( C_T > 1 \)).

7. References


Zege E.P., and A.A. Kokhanovskii, Approximate Formula for the Albedo of snow surface, Izvestiya. Atmospheric and Oceanic Physics, 33, 667-668, 1997