

The ultraviolet spectral albedo of planet earth

By JOHN E. FREDERICK, *Department of the Geophysical Sciences, The University of Chicago, 5734 South Ellis Avenue, Chicago, Illinois 60637, USA* and GEORGE N. SERAFINO, *Applied Research Corporation, 8201 Corporate Drive, Landover, Maryland 20785, USA*

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ABSTRACT

The solar backscattered ultraviolet spectral radiometer on the Nimbus 7 satellite provides a unique set of radiation measurements which allows an evaluation of the spectral albedo of the earth and its atmosphere in the wavelength range 300 to 340 nm. Near 340 nm, the derived spectral albedo expressed as a function of latitude and month exceeds that in the visible part of the spectrum, with values near 45% existing equatorward of 30° and an increase to 60%–80% toward the poles. At middle to high latitudes, the monthly mean spectral albedo exceeds the contribution from Rayleigh scattering alone by factors of 1.4 to 2.2. At wavelengths from 300 to 310 nm, where absorption by stratospheric ozone is significant, the daylight averaged spectral albedos receive negligible contribution from scattering by tropospheric clouds, yet the derived values exceed those predicted for Rayleigh scattering from a clean stratosphere. These observations are consistent with the presence of an atmospheric scattering layer, distinct from cloudiness, located at an altitude above the tropopause.

1. Introduction

The spectral albedo of the earth and its atmosphere represents a summation over an ensemble of conditions including clear and cloudy skies, and a variety of underlying surfaces. A number of researchers have viewed the albedo in a globally and spectrally integrated sense and derived a single number to characterize the entire system (e.g. Fritz, 1949; Ångström, 1962). More recent studies of the Earth's radiation budget have used global scale satellite measurements to examine the spectrally integrated albedo as a function of latitude and month (Raschke and Bandeen, 1970; Jacobowitz et al., 1984). A common element of much previous work is the emphasis on broadband visible radiation. Yet, major international concern now focuses on the biologically relevant near ultraviolet spectrum with emphasis on the spectral distribution of UV-B (280–320 nm) and UV-A (320–400 nm) radiation that reaches the ground (Worrest and Caldwell, 1986). From the standpoint of atmospheric science, one can develop the broader concept of a global budget for the biologically significant ultraviolet

radiation, analogous to the budget of visible sunlight. Major elements of an ultraviolet radiation budget are (1) the fraction of incoming UV-B and UV-A energy absorbed in the atmosphere, (2) the fraction that reaches the earth's surface, and (3) the fraction reflected back to space. This paper considers the last of these components, the spectral albedo of the earth and atmosphere in the near ultraviolet.

We define the spectral albedo to be the wavelength dependent albedo including all reflections from the solid or liquid earth plus scattering by the atmosphere containing clouds, dust, and all other components that influence the radiation field. We view this quantity, averaged over local time and a variety of surface conditions, as a function of latitude and month. In principle, the spectral albedo could be computed from an application of radiative transfer theory (e.g., Dave, 1978; Dave and Mateer, 1967) given knowledge of the ozone, cloud, dust, pressure, and surface reflectivity distributions on a global scale as functions of time. Clearly, such a calculation would be exceedingly tedious even if the required input information were available.

Fortunately, a more straightforward approach to the albedo study is possible using radiation measurements made by the solar backscattered ultraviolet (SBUV) spectral radiometer carried on the Nimbus 7 satellite (Heath et al., 1975). The primary mission of the SBUV instrument is to map the global distribution of atmospheric ozone. However, the quantities directly measured by SBUV are the extra-terrestrial solar irradiance and the radiance backscattered from the atmosphere which exists to space in the vertical direction. The data are therefore closely related to the ultraviolet spectral albedo of the earth and its atmosphere. The backscattered radiance and solar irradiance data of interest in this work are those at wavelengths, λ , of 301.9, 305.8, 312.5, and 317.5 nm in the UV-B, and 331.2 and 339.8 nm in the UV-A, all measured with a 1 nm resolution. These measured radiances implicitly contain the effects of all Rayleigh, particulate, and cloud scattering as well as absorption by ozone. The next section presents a sample of the radiation data obtained by the Nimbus 7 SBUV instrument, followed by a description of the mathematical technique used to infer daytime averaged albedos from the nadir radiance and solar irradiance measurements.

2. The radiation measurements

Inputs to the spectral albedo analysis are the ratios of the backscattered vertical radiance, $L(\lambda)$, to the incoming solar irradiance, $F(\lambda)$, where SBUV measures both of these quantities at the six wavelengths given above. Frederick and Serafino (1985) and Frederick et al. (1986) present further details of the SBUV measurement technique including the method used to derive absolute values of $L(\lambda)$ and $F(\lambda)$. Note that the "backscatter ratio", $L(\lambda)/F(\lambda)$ in the units steradian⁻¹, is not itself the spectral albedo since this latter quantity involves integration over the angular distribution of radiance emerging from the upper hemisphere. The SBUV measurement gives only the vertical component of this radiation field.

At wavelengths shorter than 295 nm the emergent radiance overwhelmingly represents Rayleigh scattering modified by ozone absorption. However, longward of 300–310 nm the

incoming solar irradiance penetrates in part to the troposphere and to the Earth's surface. In this case, scattering by clouds or dust and reflection from the solid or liquid lower boundary become significant factors. This paper refers to these processes taken collectively as "non-Rayleigh scattering" to distinguish them from molecular scattering. One of the objectives of this work is to evaluate the non-Rayleigh scattering contribution to the spectral albedo of the earth and its atmosphere.

Figs. 1 through 3 present backscatter ratios, $L(\lambda)/F(\lambda)$, at wavelengths 301.9, 317.5, and 339.8 nm plotted against solar zenith angle (SZA). Values refer to the latitude bin 0–10°N for the month of March 1979. The Nimbus 7 satellite is in a sun synchronous orbit at a local time of noon in the daylight hemisphere. Hence, in a single month and latitude band the data cover a limited range of SZA. Fig. 1 for 301.9 nm shows a spread in backscatter ratios of approximately a factor of two at any SZA, although there is a tendency for the points to cluster near the smaller values indicative of clear sky conditions. The backscatter ratios at 317.5 nm in Fig. 2 are an order of magnitude larger than at 301.9 nm, and the width of the data envelope is greater. This is as expected, since the ozone absorption cross section decreases by a nearly a factor of 10 between 310.9 and 317.5 nm. Thus, a much greater fraction of the incoming radiation at 317.5 nm reaches the lower boundary of the

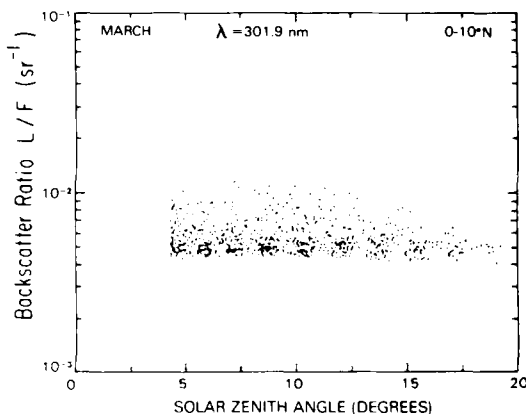


Fig. 1. Backscatter ratios L/F (units sr^{-1}) measured at a wavelength of 301.9 nm during March 1979 in the latitude band 0–10°N.

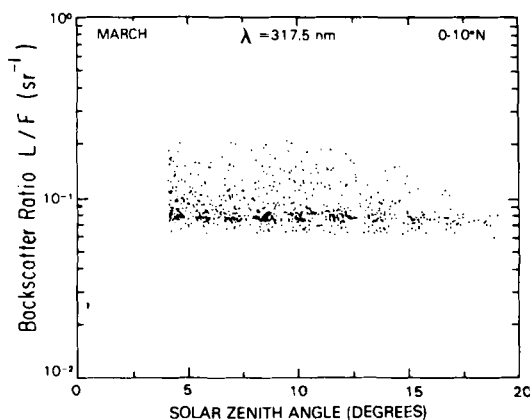


Fig. 2. Backscatter ratios L/F (units sr^{-1}) measured at a wavelength of 317.5 nm during March 1979 in the latitude band 0–10°N.

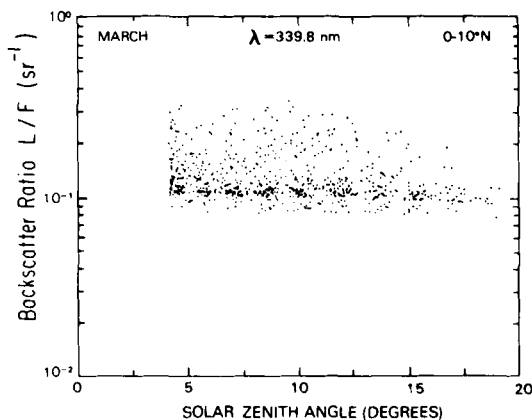


Fig. 3. Backscatter ratios L/F (units sr^{-1}) measured at a wavelength of 339.8 nm during March 1979 in the latitude band 0–10°N.

atmosphere than is the case at 301.9 nm. The reflectivity of this boundary is highly variable depending on cloudcover and surface type, and this leads to the variations observed in backscattered radiance. The behaviour at 339.8 nm shown in Fig. 3 is similar to that at 317.5 nm, although the optical depth for absorption has shrunk by another order of magnitude. The mean ozone column abundance for 0–10°N in March derived from SBUV, 0.25 cm, gives a vertical optical depth for absorption on the order of 10^{-1} at 339.8 nm. Therefore, the backscatter ratios at 339.8 nm are determined almost entirely by scattering and surface reflection. The measured values now vary over a factor of three at most SZAs.

Figs. 4 through 6 are analogous to those presented above except that the latitude band is now 40–50°N. Both the slant path taken by sunlight through the atmosphere and the column ozone abundance, being 0.39 cm, are greater here than in the tropics. Therefore, at a wavelength where absorption by ozone is strong, a negligible fraction of the incoming solar irradiance reaches the troposphere, is reflected, and emerges to space. The observed signal here consists entirely of Rayleigh scattering taking place above the tropopause. The 301.9 nm signal in Fig. 4 is consistent with this reasoning. A comparison with Fig. 1 reveals both smaller backscatter ratios as well as a greatly reduced scatter at middle latitudes. However, at wavelengths where the

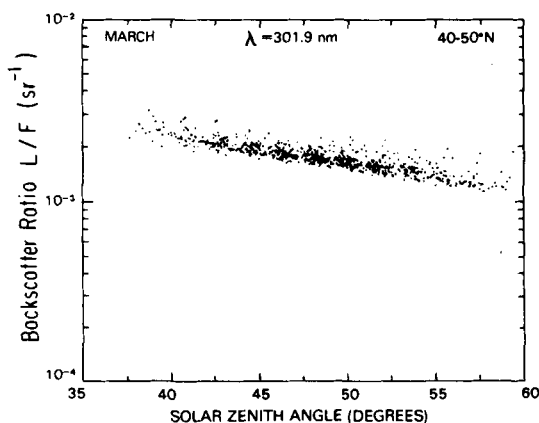


Fig. 4. Backscatter ratios L/F (units sr^{-1}) measured at a wavelength of 301.9 nm during March 1979 in the latitude band 40–50°N.

optical depth is less, as in Figs. 5 and 6 for 317.5 and 339.8 nm, respectively, variable cloud and surface conditions continue to produce large fluctuations in the radiance measurements.

Two significant points are evident in Fig. 1 through 6. First, the backscatter ratio, and therefore the spectral albedo, varies rapidly with wavelength in the biologically active ultraviolet region. Second, the concept of a global albedo in the UV-B and UV-A must represent a mean over an ensemble of values that individually vary by a

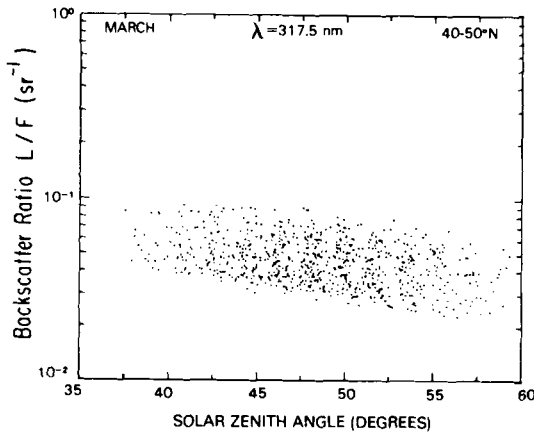


Fig. 5. Backscatter ratios L/F (units sr^{-1}) measured at a wavelength of 317.5 nm during March 1979 in the latitude band 40–50°N.

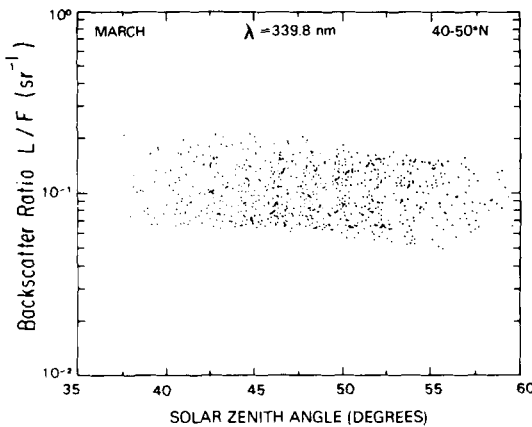


Fig. 6. Backscatter ratios L/F (units sr^{-1}) measured at a wavelength of 339.8 nm during March 1979 in the latitude band 40–50°N.

factor of two or more at a given location over the course of a month.

3. Mathematical formulation

The relevant quantity for this study is the ratio of the flux that escapes the earth's atmosphere at a given wavelength, integrated over the daylight portion of a 24-hour period, to the corresponding integrated incoming flux falling on a horizontal surface. This daytime averaged spectral albedo at

a given location is defined as:

$$A = \frac{\int dt F_{\text{out}}(\lambda, t)}{F(\lambda) \int dt \mu_o(t)}, \quad (1)$$

where the integrals over time t cover the daylight portion of a 24-hour period, μ_o is the cosine of the time dependent solar zenith angle, $F(\lambda)$ is the extraterrestrial solar irradiance measured by SBUV, and $F_{\text{out}}(\lambda, t)$ is the outgoing flux which escapes to space. The objective of the mathematical development is to recast (1) in a form that allows direct use of the SBUV measurable, $L(\lambda)/F(\lambda)$. Jacobowitz et al. (1984) and Arking and Vemury (1984) have discussed analogous procedures for use with data from the Nimbus 7 earth radiation budget experiment.

The outgoing flux is related to the emergent radiance by:

$$F_{\text{out}}(\lambda, t) = \int d\phi' \int d\mu' \mu' L_{\text{out}}(\lambda, t, \mu', \phi'), \quad (2)$$

where μ' is the cosine of the angle from the vertical and ϕ' is the azimuthal angle relative to the position of the sun. The measured SBUV radiance is the vertical component, $L(\lambda) \equiv L_{\text{out}}(\lambda, t_o, \mu' = 0)$, at the fixed local time, t_o , of the Nimbus 7 orbit. The daytime averaged albedo is then:

$$A = \frac{L(\lambda)}{F(\lambda)} \left\{ \frac{\int dt \int d\phi' \int d\mu' \mu' L_{\text{out}}(\lambda, t, \mu', \phi') / L(\lambda)}{\int dt \mu_o(t)} \right\} \quad (3)$$

or:

$$A = G(\lambda) \frac{L(\lambda)}{F(\lambda)}, \quad (4)$$

where $G(\lambda)$, in steradians, is defined as the bracketed term in eq. (3), and the quantity containing $L(\lambda)$ implicit in G is now computed from the same radiation model used to generate L_{out} at all emergent angles and local times. Use of the SBUV observable, $L(\lambda)/F(\lambda)$, in eq. 4 implicitly incorporates scattering by clouds and all other processes that influence the radiation field into the results.

The quantity $G(\lambda)$ is a scaling factor that accounts for the angular distribution of radiation emerging over the upper hemisphere and for local time variations, both in the extraterrestrial flux incident on a horizontal surface and in the flux returned to space. For each set of $L(\lambda)/F(\lambda)$ ratios

at the six wavelengths analyzed here, the SBUV data set includes corresponding values of the solar zenith angle, atmospheric ozone, and effective Lambertian surface reflectivity derived from the measurements (Klenk et al., 1982). To compute the appropriate $G(\lambda)$ value for each $L(\lambda)/F(\lambda)$ measurement, we performed a radiation transfer calculation including all orders of multiple scattering, absorption by ozone, and reflection from a lower boundary. The numerical technique is that discussed by Herman and Browning (1965). To verify proper operation we input the ozone and surface reflectivity results provided with the SBUV data set to compute the corresponding $L(\lambda)/F(\lambda)$ values. Note that these geophysical data products were derived from the SBUV radiation measurements as described by Klenk et al. (1982) using algorithms that are totally independent of our own. We found consistent agreement between the measured backscatter ratios and those produced by our calculation to better than 1%. We then applied the model to predict the angular distribution of emergent radiation and its local time dependence. In practice, we generated a grid of $G(\lambda)$ values as a function of wavelength, ozone amount, lower boundary reflectivity, solar zenith angle, latitude, and month. This allowed rapid derivation of the $G(\lambda)$ value appropriate to a given $L(\lambda)/F(\lambda)$ measurement.

The procedure described above assumes the angular distribution and local time dependence of radiation emerging from the atmosphere to be identical to that predicted by our calculations. In the presence of clouds, an error could result for the following reason. A cloud appears in the radiative transfer model as a Lambertian lower boundary of high reflectivity placed at a climatological cloudtop pressure. To the extent that the Lambertian assumption is invalid, the accuracy of the computed angular distribution of emergent radiation degrades. However, cloud configurations are highly variable, and the error encountered in any one albedo determination may be positive or negative. If the errors incurred in treating cloud reflection are random, they will tend to cancel in computing the mean albedos, which include over 1000 measurements in each 10 degree wide latitude band and month. At present we cannot quantify the uncertainties in this procedure further.

4. Results

Before we present spectral albedos derived from the satellite-based radiation measurements, it is useful to restate the variables on which the measured signal depends. The backscattered radiance consists of both a Rayleigh and non-Rayleigh scattered component. As wavelength decreases and optical depth increases, a greater fraction of the observed signal originates from above the tropopause, and the measurement therefore carries less information on surface and cloud reflectivity. In the absence of absorbers, an increase in solar zenith angle leads to an increase in spectral albedo both because of the μ_0 factor in the denominator of eq. (1) and the fact that the increased slant path column density traversed by the incoming beam promotes enhanced backscattering. However, the addition of an absorbing gas complicates the scenario. The solar zenith angle variation of the spectral albedo now depends on the relative efficiencies of absorption and scattering over the range of altitude which contributes to the signal that emerges to space. These solar zenith angle variations are implicit in the local time integration of eq. (1) and appear in the daytime averaged albedos as a dependence on latitude and month.

Fig. 7 illustrates the points made above by presenting the spectral albedo as a function of wavelength for the latitude bands 0–10°N and 40–50°N during March. In general, mean values reported for each 10 degree wide latitude bin in Fig. 7 and subsequent figures are based on 1200 to 1700 individual $L(\lambda)/F(\lambda)$ measurements at each of six wavelengths during a given month. The scatter among individual data points is as shown in Fig. 1 through 6. The prominent feature of Fig. 7 is the large increase in albedo with wavelength which arises from the decreasing strength of absorption by ozone. When this absorption is relatively weak, as at 331.2 and 339.8 nm, the larger SZAs at 40–50°N imply a larger spectral albedo here than at 0–10°N. In the UV-B the SZA dependence of Rayleigh scattering still promotes large albedos at 40–50°N. However, the increase in ozone from 0.25 cm at 0–10°N to 0.39 cm at 40–50°N combined with the increased slant path of sunlight through the absorber leads to a decrease in spectral albedo with the increase in latitude.

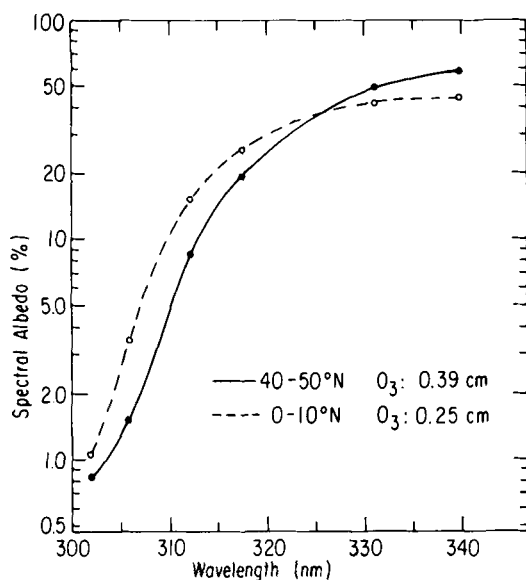


Fig. 7. Spectral albedo expressed as a function of wavelength for March 1979 in the latitude bands 0–10°N and 40–50°N. The total column ozone amounts are 0.25 and 0.39 cm respectively. Dots mark the wavelengths at which SBUV obtains measurements.

We now extend the analysis to the global scale and examine both daytime mean albedos and the physical mechanisms responsible for geographic and temporal variations. Figs. 8 through 11 present the latitudinal and monthly behavior in spectral albedo at 339.8, 317.5, 305.8, and 301.9 nm respectively. Within 20° to 30° of the equator the 339.8 nm spectral albedo remains near 45%. Values increase toward the poles in all seasons, reaching 80% in the 60° to 80° latitude zone of both hemispheres. The geographic patterns in Fig. 8 are similar to those presented by Jacobowitz et al. (1984) for broadland measurements from the earth radiation budget experiment. However, the spectral albedos at 339.8 nm exceed those deduced in the visible because of the increased efficiency of Rayleigh scattering with decreased wavelengths. A global and annual mean result at 339.8 nm is near 50% as compared to 30–38% in the visible part of the spectrum (Ångström, 1962; SMIC, 1971).

Figs. 9–11 show latitudinal and seasonal patterns that vary with wavelength. Fig. 9, for 317.5 nm, reveals the increasing impact of ozone on the

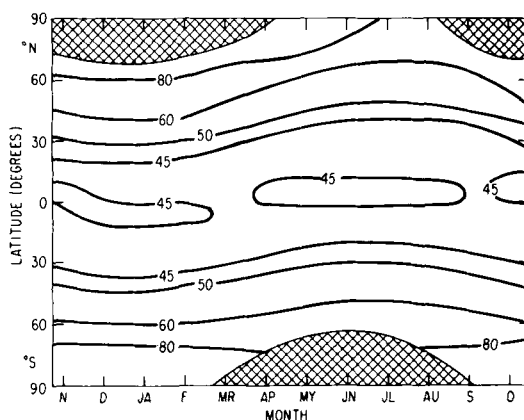


Fig. 8. Distribution of daytime averaged spectral albedo as a function of latitude and month over the period November 1978 through October 1979. The wavelength is 339.8 nm. Contour labels are in percent. Darkened areas indicate regions where measurements are not available.

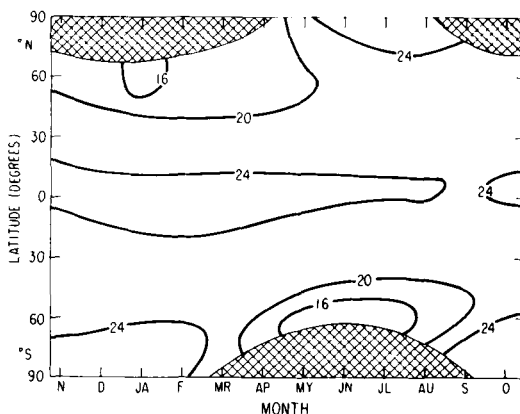


Fig. 9. Distribution of daytime averaged spectral albedo as a function of latitude and month over the period November 1978 through October 1979. The wavelength is 317.5 nm. Contour labels are in percent. Darkened areas indicate regions where measurements are not available.

spectral albedo, with values between 16% and 24%. In the period from mid-autumn through late spring the spectral albedo decreases with increasing latitude in both hemispheres, and the largest gradients exist near the solstices. Spectral albedos reach minima in summer between latitudes 30° and 50° in both hemispheres and in-

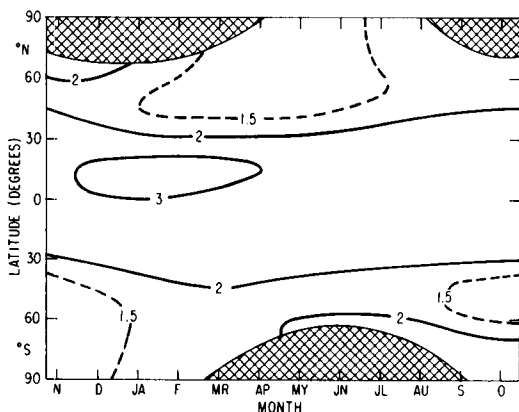


Fig. 10 Distribution of daytime averaged spectral albedo as a function of latitude and month over the period November 1978 through December 1979. The wavelength is 305.8 nm. Contour labels are in percent. Darkened areas indicate regions where measurements are not available.

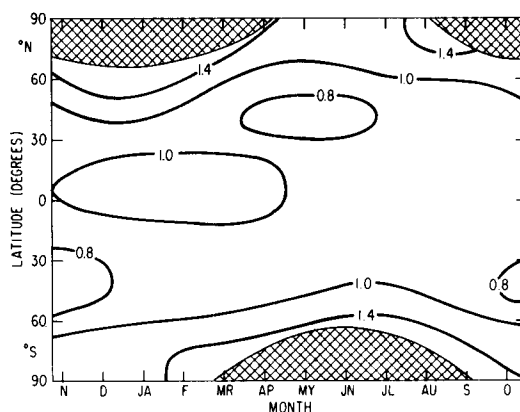


Fig. 11. Distribution of daytime averaged spectral albedo as a function of latitude and month over the period November 1978 through October 1979. The wavelength is 301.9 nm. Contour labels are in percent. Darkened areas indicate regions where measurements are not available.

crease to 24% toward the poles. At 305.8 nm absorption by ozone limits the spectral albedo to 1.5%–3.0%. Results tend to decrease with increasing latitude in both hemispheres, although values in the 2.0%–2.5% range exist poleward of 60° from May through the spring. The spectral albedo at 301.9 nm shown in Fig. 11 lies between 0.8% and 1.0% over most of the globe equatorward of 60° latitude, while values increase to 1.4% in autumn and winter.

A set of radiative transfer calculations allows us to separate the derived spectral albedos into a component that originates from Rayleigh scattering at altitudes above the climatological cloudtop location and a component that arises from altitudes at and below the cloudtops, including reflection from the ground, scattering from atmospheric molecules located between the cloud and the ground, and scattering from the clouds themselves. For ease of discussion, we refer to the latter component as the “non-Rayleigh” contribution even though a portion of this arises from molecular scattering beneath clouds. The wavelength dependencies of the ozone absorption and Rayleigh scattering cross sections require radiation at progressively longer wavelengths to penetrate deeper into the atmosphere prior to scattering. We here demonstrate this by the sample calculations shown in Fig. 12. These results

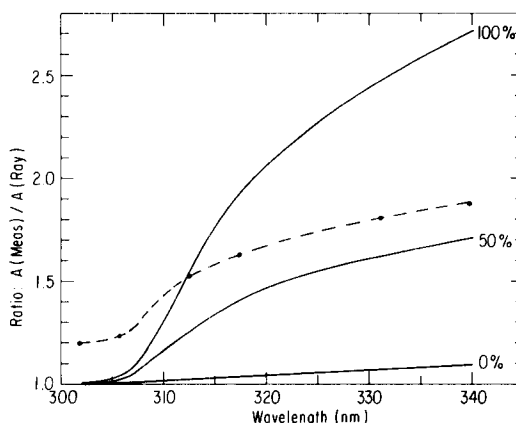


Fig. 12. Solid curves give the computed effect of reflection from clouds and the ground on the spectral albedo. The curves refer to skies which are 100%, 50%, and 0% cloudcovered. Dashed curve gives results based on spectral albedos derived from SBUV measurements for 40–50°N in April. See text for details of calculations.

illustrate the physical mechanisms which determine the spectral albedos derived from SBUV. However, we do not seek to duplicate the albedos of Figs. 8–11 in a rigorous fashion as this would require a much more sophisticated treatment of non-Rayleigh scattering than we have available

to us. To produce Fig. 12, we computed the clear sky spectral albedo at two hour intervals during the course of a day using the monthly mean ozone derived for 40°–50°N in April. We then averaged these results to obtain a daytime mean value analogous to those derived from the satellite $L(\lambda)/F(\lambda)$ data. We performed these calculations for surface reflectivities of 0.1 and 0.0, where the former is a substitute for results derived from SBUV, $A(\text{Meas})$, and the latter represents the Rayleigh scattering contribution, $A(\text{Ray})$. We then repeated these calculations with the lower boundary of the atmosphere at 6.5 km to simulate a cloudtop. Here a surface reflectivity of 0.65 produced synthetic spectral albedos, $A(\text{Meas})$, in the presence of 100% cloudcover, while a reflectivity of 0.0 gave the Rayleigh scattering contribution, $A(\text{Ray})$.

The ratios $N = A(\text{Meas})/A(\text{Ray})$ for each wavelength are a measure of the non-Rayleigh contribution to the spectral albedo. Fig. 12 presents theoretical N -values versus wavelength for clear sky and 100% cloudcovered conditions based on the calculations discussed above. We also simulated a fractional cloudcover of 50% by combining albedos computed for clear and cloudy conditions. Fig. 12 demonstrates that daytime averaged spectral albedos at wavelengths less than 307 nm are insensitive to cloudcover located at tropospheric altitudes. The spectral albedo computed for 305.8 nm is within 5% of the pure Rayleigh scattering value, while that for 301.9 nm is independent of surface and cloudcover conditions when averaged over the day. Spectral albedos at longer wavelengths show a substantial response to cloud reflection, and under 100% cloudcovered conditions the non-Rayleigh contribution exceeds that from Rayleigh scattering at all wavelengths in the UV-A.

Separation of the derived spectral albedo into its Rayleigh and non-Rayleigh components requires application of the radiative transfer model referenced previously. For each individual spectral albedo that entered the means of Figs. 8 through 11, we estimated the fractional cloudcover in the SBUV field of view using the derived Lambertian reflectivity and relationships presented by Klenk et al. (1982). Using the global distribution of ozone measured by the SBUV instrument and a reflectivity of zero at the ground or cloudtop pressure, we computed the latitudinal

and seasonal distribution of albedo arising from Rayleigh scattering alone. The N -value derived for each 10° latitude band and month is the ratio of the mean albedo deduced from SBUV to the mean Rayleigh scattering contribution. Figs. 13 and 14 present the N -values derived for wavelengths 339.8 and 305.8 nm, respectively. The non-Rayleigh contribution to the 339.8 nm spectral albedo is compatible with the illustrative results of Fig. 12. The contribution from clouds, the sub-cloud atmospheric layer, and the ground increases the spectral albedo to 1.4 to 2.2 times

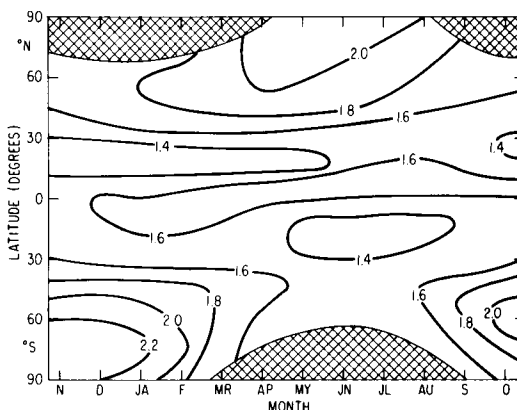


Fig. 13. Ratios of the spectral albedos deduced for a wavelength of 339.8 nm to those computed for Rayleigh scattering alone over the period November 1978 to October 1979. Darkened areas indicate regions where measurements are not available.

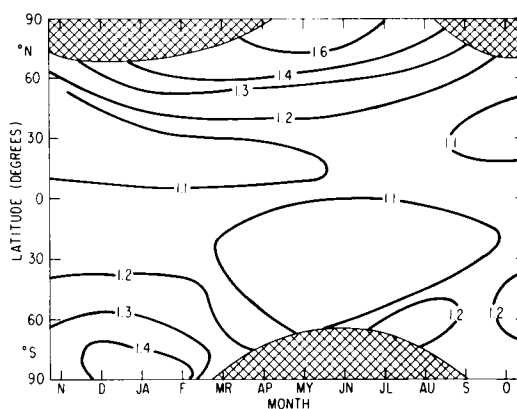


Fig. 14. Ratios of the spectral albedos deduced for a wavelength of 305.8 nm to those computed for Rayleigh scattering alone over the period November 1978 to October 1979. Darkened areas indicate regions where measurements are not available.

the Rayleigh scattering component alone. The ratios of Fig. 14 for 305.8 nm show a geographic pattern similar to that at 339.8 nm, although the shorter wavelength results are consistently smaller. Spectral albedos at 305.8 nm are 1.1–1.6 times the Rayleigh scattering contributions.

Fig. 12 includes N -values based on spectral albedos derived from SBUV for 40–50°N in April. At 301.9 nm and 305.8 nm, the values exceed those predicted even for 100% cloudcover, while the 312.5 nm result is larger than one can accept for realistic cloudcover conditions. Yet, the theoretical curves in Fig. 12 show that it is not possible for reflecting surfaces placed in the troposphere to lead to the observed albedos. We also performed sample calculations that placed the reflecting boundary of the atmosphere at 10 km in altitude and found negligible change from the results in Fig. 12 for the two shortest wavelengths. The conclusion is that the non-Rayleigh component of spectral albedo at 301.9 and 305.8 nm originates at altitudes above the tropopause.

Consider a scattering region, composed of aerosol, located at a sufficiently high altitude that a significant fraction of the atmospheric ozone column lies beneath it. A portion of the incoming radiation at 331.2 and 339.8 nm will be scattered from this layer. However, a probable fate for photons that penetrate through the layer is Rayleigh backscattering or reflection from clouds. In either case, the result is a signal that emerges to space, and therefore the high altitude scatterers have a relatively minor influence on the spectral albedo at these wavelengths. However, when absorption by ozone becomes significant, as at 301.9 and 305.8 nm, scattering by the high altitude layer prevents penetration into the lower region where ozone would further deplete the signal. Therefore, a significant enhancement in the spectral albedo results at these wavelengths. Note that the growing importance of the non-Rayleigh component with latitude shown in Fig. 14 does not necessarily imply an increase in optical thickness of the scattering layer. The

increasing angle of the incoming solar beam alone could produce the observed behaviour. The altitude of the scattering region is not well-defined. However, since the contribution functions for the 301.9 and 305.8 nm signals maximize in the low to middle stratosphere (Mateer, 1977), this is a likely location of the layer. The background stratospheric aerosol layer (Russell et al., 1981a,b) is a potential source of the albedo enhancement, although definitive statements are not possible at present.

5. Concluding remarks

The spectral albedo of the earth in the near ultraviolet represents the combined action of Rayleigh scattering, scattering from clouds and aerosols, and absorption by ozone. At any point on the globe, the instantaneous spectral albedo is highly variable with local conditions, and the mean value for a given latitude zone and month is significantly greater than would exist for a pure Rayleigh scattering atmosphere. At 339.8 nm, the true spectral albedo exceeds the contribution from Rayleigh scattering alone by 40%–120%. At wavelengths where absorption by ozone is significant, the zonally and monthly averaged spectral albedo receives significant contributions from a component other than Rayleigh scattering. This enhancement is most pronounced at high latitudes, and poleward of 60° the spectral albedos are greater than expected for Rayleigh scattering alone by factors of 1.3 to 1.6. Results at wavelengths 301.9 and 305.8 nm suggest that essentially all of the non-Rayleigh scattering at these wavelengths takes place at altitudes above the tropopause.

6. Acknowledgement

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