ENVIRONMENTAL SCIENCES 1999 Vol. 3, No. 1. 41~56
Bulletin of the Korean Environmental Sciences Society

Parameterization Model for Damaging Ultraviolet-B Irradiance

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(Manuscript received on 5 September 1998)

Since UV-B radiation measuring networks have not been established, numerical models which calculate the flux from other readily available meteorological measurements may play an important role. That is, such a problem can be solved by using parameterization models such as two stream approximation, the delta-Eddington method, doubling method, and discrete ordinate method. However, most UV-B radiative transfer models have not been validated with measurements, because such models are not intended as practical computational schemes for providing surface estimates of UV-B radiation. The main concern so far has been to demonstrate model sensitivity for cloudless skies. In particular, few have been concerned with real cloud information. Clouds and aerosols have generally been incorporated as constituents of particular atmospheric layers with specified optical depths and scattering properties.

The parameterization model presented here is a combination of a detailed radiative transfer algorithm for a cloudless sky radiative process and a more approximate scheme to handle cloud effects. The model input data requires a daily measurement of the total ozone amount plus a daily record of the amount and type of cloud in the atmosphere. Measurements for an examination of the models at the Department of Atmospheric Sciences, Pusan National University have been taken from February, 1995. These models can be used to calculate present and future fluxes where measurements have not been taken, and construct climatologies for the period before ozone depletion began.

Key word : UV-B radiation, parameterization model

1. Introduction

The possible increase of biologically damaging UV-B irradiance due to the depletion of the ozone plus changes in the important photochemical reaction in the troposphere have become of great concern in the world.[1] For example, a 1% reduction in ozone is expected to yield a 2-3% increase in the incidence of skin cancer.[2] Extensive theoretical studies have been conducted on the increase of UV-B due to ozone depletion since the 1970's[3,4,5,6], however, so far there have been no observational studies.[7]

Since UV-B radiation measuring networks have not been established, parameterization and numerical models which can calculate flux from other readily available meteorological measurements may play an important role. Such problems can be solved by using radiative transfer models such as two stream approximation, the delta-Eddington method, doubling method, and discrete ordinate method. However, most UV-B radiative transfer models have not been validated with measurements, because such models are not intended as practical computational schemes for providing surface estimates of UV-B radiation.

The main emphasis so far has been to demonstrate model sensitivity[8,9,10,11,12,13,14,15] for cloudless skies. In particular, few have been concerned with real cloud information.[16] Clouds have generally been incorporated as constituents of particular atmospheric layers with specified optical depths and scattering properties.[10,17,18,19,20] Some UV-B variation is regular and due to well-determined geometric factors such as the position of the sun in the sky(;solar zenith angle) and the yearly alteration in the earth-sun distance. Other variations are due to changes in atmospheric constituents which determine the transmission of the radiation from the top of the atmosphere to the surface; clouds, ozone, and aerosols are the most important among these, yet minor trace gases(e.g., nitrogen dioxide and sulfur dioxide) can also contribute in highly polluted areas.[21]

Accordingly, a parameterization model of the spectral integration of the damaging UV-B band, a solar radiation spectrum from 280 nm to 320 nm, can deduce the relation between the global, direct, and diffuse radiation flux under cloudless skies and the influences of solar elevation, total ozone content, albedo, turbidity, and altitude on the UV-B radiation flux. Furthermore, the influence of cloudiness can also be analyzed. The model input data requires a daily measurement of the total ozone amount and a daily record of cloud amount and type at different levels in the atmosphere. This model can then be used to calculate present and future fluxes where UV-B measurements have not been made.

2. Data

At Pusan National University(PNU; 35.23°N, 129.07°E), the observation of global solar UV-B irradiance using a UV-Biometer started from 26 February, 1996. The UV-Biometer that is connected to the worldwide network for UV-B monitoring was made sensitive to the wavelength range of UV-B by a combination of special filters, plus it includes a spectral sensitivity adapted to the action spectrum of the biological reaction of erythema in human skin.[22,23,24] The measurement of the UV-B MED(Minimum Erythema Dose: 1 MED = 21 mJ/cm^2) within the biologically harmful ultraviolet region of 280 nm to 320 nm was taken by intergrating every 5 minutes from sunrise to sunset every day. The daily TOMS(Total Mapping Spectrometer) and ozone sonde(Pohang Meteorological Agency; measurements of the total ozone amount in the atmospheric column) data were used to adjust the selected parameterization model. Cloud and visibility data from the Pusan Meteorological Agency (PMA) were also used.

3. Parameterization model

The parameterization model is adapted from the original model of Davies and Hay.[25] The estimation of solar radiation is dictated by the atmosphere which absorbs and scatters radiation. To account for these processes the radiative transfer equation as presented by Kondrat'yev[26] is used

$$\frac{\mu dI_\lambda(\tau_\lambda,\mu,\phi)}{d\tau_\lambda} = -I_\lambda(\tau_\lambda,\mu,\phi) + \frac{\omega_0}{4\pi}\int_0^{2\pi}\int_0^1 I_\lambda(\tau_\lambda,\mu,\phi)p(\tau_\lambda,\mu',\phi'\rightarrow\mu,\phi)d\mu' d\phi' \quad (1)$$

where I_λ is the spectral radiation intensity at wavelength λ, μ is the cosine of the solar zenith angle θ, τ_λ is the atmospheric optical depth, $p(\tau_\lambda, y', y)$ is the phase function which accounts for the scattering of an incident ray from direction μ', ϕ' into direction μ, ϕ. ω_0 is the single scattering albedo, and ω is the solid angle.

If x_λ and σ_λ are the mass absorption and scattering coefficients, respectively, the optical depth and single scattering albedo, which is the ratio of energy scattered by aerosols to total attenuation under the first impingement by direct radiation are given by

$$\tau_\lambda = \int_0^z (x_\lambda + \sigma_\lambda) dz \qquad (2)$$

$$\omega_0 = \frac{\sigma_\lambda}{x_\lambda + \sigma_\lambda} \qquad (3)$$

where z is the height of the aerosol atmosphere. The Phase function can be usefully characterized by the asymmetry factor g_λ which is defined by

$$g_\lambda = \langle \cos\psi \rangle = \int_{-1}^{+1} P_\lambda(\cos\psi)\cos\psi d\cos\psi \qquad (4)$$

where ψ is the scattering angle. Physically g is the difference between the flux densities in the forward direction and backward direction arising from the scattering by a particle when the incident flux is normalized to 1.[27]

3.1. Surface UV-B irradiance under cloudless skies

The total cloudless sky irradiance at surface G_0 is the sum of the contributing components.

$$G_0 = I_0 + D_R + D_A + D_S \qquad (5)$$

where I_0 is the direct beam irradiance for cloudless skies, D_R is the diffuse irradiance due to Rayleigh scattering. D_A is the diffuse irradiance due to aerosol scattering, and D_S is the diffuse irradiance due to multiple reflections.

The attenuations are shown schematically in Fig. 1.

3.1.1. Direct beam irradiance for cloudless skies

For direct beam radiation the phase function from the radiative transfer equation can be ignored. Thus, equation (1) is simplified and can be solved. Integrating between the ground and the top of the atmosphere, z, the direct beam radiation received on the ground at normal incidence is given by Beer's law

$$\frac{I_\lambda(\tau)}{I_\lambda(0)} = \exp\left(\frac{\tau_\lambda}{\mu}\right) \qquad (6)$$

The τ_λ term includes all of the principal absorbers and scatters responsible for the attenuation. Thus,

$$\tau = \tau_{o3\lambda} + \tau_{R\lambda} + \tau_{a\lambda} \qquad (7)$$

where $\tau_{o3\lambda}$, $\tau_{R\lambda}$, $\tau_{a\lambda}$ are the spectral optical depths of a direct beam due to absorption by ozone(O_3), Rayleigh molecular scattering, and extinction(scattering and absorption) by aerosols respectively.

An exponential function defines the total transmittance which has several component terms. The monochromatic distribution of direct beam radiation transmitted through a cloudless atmosphere and received on a horizontal surface is

$$I_0 = I(0)\cos\theta\,[\tau_{03}\ \tau_R\ \tau_a\ ] \qquad (8)$$

$$I(0) = I(0\ \lambda)(R_0/R)^2 \qquad (9)$$

where τ_{03} is the transmissions after absorption by ozone, τ_R is the transmissions after absorption by Rayleigh scattering, τ_a is the transmissions after absorption and scattering by aerosol, θ is the zenith angle. R_0 is the mean sun-earth

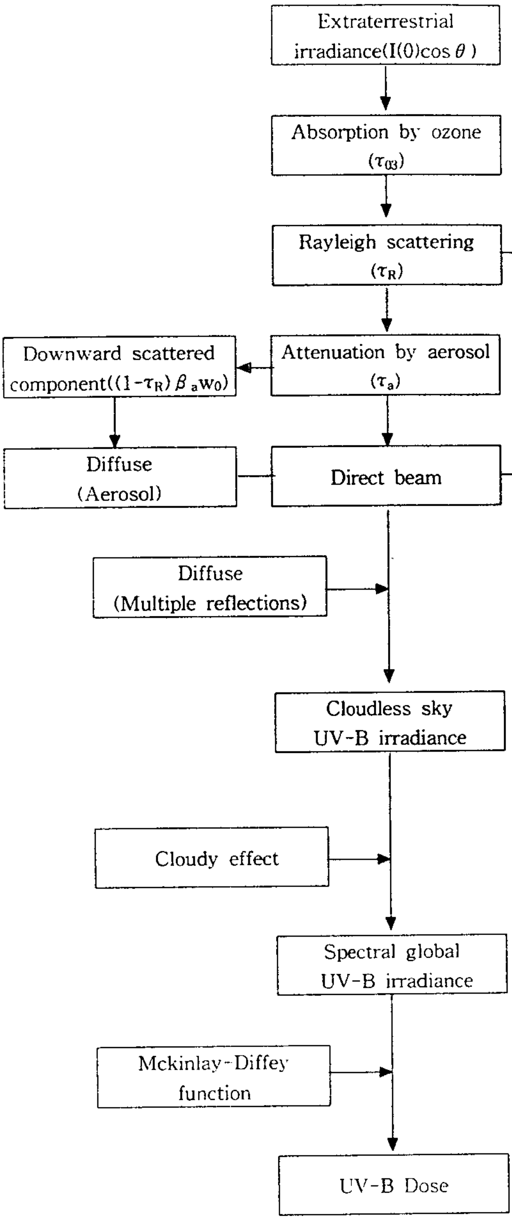

Fig. 1. Flow chart of model for calculating UV-B MED.

distance(=1.496×10^8 km), R is the sun-earth distance for any day of any year, I(0) is the corrected solar constant, and I(0 λ) is the spectral solar constant.[28)]

Traditionally, the distance r between the Earth and the Sun is expressed in terms of a Fourier series type of expansion with a number of coefficients. With a maximum error of 0.0001. Spencer[29)] developed the following expression for the reciprocal of the square of the radius vector of the earth, here called the eccentricity correction factor of the earth's orbit

$$(R_0 / R)^2 = 1.00011 + 0.034221\cos\theta_0 + 0.00128\sin\theta_0 - 0.000719\cos2\theta_0 + 0.000077\sin2\theta_0 \quad (10)$$

$$\theta_0 = 2\pi(d_n - 1)/365 \quad (11)$$

where θ_0 is the angle (in radians) defined by the day number and d_n is day number range from 0 on January 1 to 365 on December 31. February is always assumed to have 28 days.

With ultraviolet, the results by Vigroux[30] are commonly employed. Vigroux presents attenuation coefficients in a form suitable for Bouguer's law. The transmission after absorption by ozone, τ_{03} is

$$\tau_{03} = \exp(-k_{03\lambda} l_{03} m_r). \quad (12)$$

where l_{03} is the vertical ozone layer thickness (cm), m_r is the relative optical air mass and $k_{03}\lambda$ is the attenuation coefficient caused by ozone (Table 1).

Calculation of relative optical air mass.[31]

Table 1. Spectral absorption coefficient $k_{o3}\lambda$ for ozone[32](Lecker).

λ (nm)	$k_{o3}\lambda$ (cm^{-1})
0.290	38.000
0.295	20.000
0.300	10.000
0.305	4.800
0.310	2.700
0.315	1.350
0.320	0.800
0.325	0.380
0.330	0.160
0.335	0.075
0.340	0.040
0.345	0.019
0.350	0.007
0.355	0.000
0.400	0.000

$$m_r = [35/(1224\mu_0^2 + 1)]p/p_0 \quad (13)$$

$$\mu_0 = \sin\varphi\sin\delta + \cos\varphi\cos\delta\cos h \quad (14)$$

$$h = 15^\circ |12 - LAT| \quad (15)$$

$$LAT = LST + ET/60 + (LMS - LS)/15 \quad (16)$$

$$\delta = 0.4093 \times \sin(0.01698 \times (d_n - 80)) \quad (17)$$

$$ET = 0.000075 + 0.001868\cos\theta_0 - 0.032077\sin\theta_0 - 0.14615\cos2\theta_0 - 0.040840\sin2\theta_0 \quad (18)$$

where p is the station pressure (kPa), p_0 is the sea level pressure (101.3 kPa), μ_0 is $\cos\theta$, φ is the station latitude, δ is the solar declination, h is the solar hour angle, LAT is the local apparent time, LST is the local standard time, ET is the equation of time, LMS is the standard meridian of the appropriate time zone, and LS is the longitude of the station

Because of the variation of the Rayleigh scattering coefficient with λ^{-4}, the spectral transmittance of air molecules rapidly increases with wavelength and decreases with an increasing optical air mass. The complete expression for transmission after absorption by Rayleigh scattering, τ_R[33] can be rewritten as

$$\tau_R = \exp(-0.008735\lambda^{-4.08} m_r). \quad (19)$$

Absorption by an aerosol may display some spectral selectivity, however, it is not expected to be highly structured. In general, scattering is much greater than absorption. Since the attenuation effects of scattering and absorption by dust are difficult to separate, Angstrom[34,35] suggests a single formula generally known as Angstrom's turbidity formula. It is called turbidity because the scattering of solar radiation by matter other than dry air molecules is called turbidity of the atmosphere(in the optical sense). Consequently, it includes attenuation due to dry as well as wet dust particles- that are all aerosols. Using the Angstrom's turbidity formula, transmission after absorption and scattering by aerosol, τ_a[33] can be written as

$$\tau_a = \exp(-\beta_A \lambda^{-\alpha w} m_r) \tag{20}$$

In this formula β_A is called Angstrom's turbidity coefficient, αw is the wavelength exponent, and the wavelength λ is in micrometers. β_A, which varies from 0,0 to 0.5 or even higher, is an index reprehension of the amount of aerosols present in the atmosphere in the vertical direction. The wavelength exponent αw is related to the size distribution of the aerosol particles. Large values of αw indicate a relatively high ratio of small particles to large particles. Generally, αw has value of between 0.5 and 2.5 : a value of 1.3 is commonly employed as this was originally suggested by Angstrom. A good average value for most natural atmospheres is $\alpha w = 1.3 \pm 0.5$.

The turbidity parameter β_A can be determined from a measurement of visibility in the horizontal direction, if such a measurement exists. This visibility is also called the meteorological range. For visibilities greater than 5 km, the turbidity parameter can be determined from the following equation developed by McClatchey and Selby[33] and others :

$$\beta_A = (0.55)^{\alpha w}[(3.912/\text{vis} - 0.01162)(0.02472(\text{vis}-5) + 1.132)] \tag{21}$$

where vis is visibility in kilometers. In order to predict β_A, it is necessary first to guess the value of αw.

3.1.2. Diffuse irradiance for cloudless skies

Diffuse radiation is generated by the scattering effects of air molecules and aerosols. Diffuse radiation is generated by the first impingement of the direct radiation, called primary scattering. This diffuse radiation in turn strikes other molecules and particles and thus the scattering process continues. This continued scattering process is called multiple scattering. An analysis of multiple scattering is mathematically very complex and requires a great deal of computational time. Furthermore, the contribution of multiple scattering is often small, and any gain from an exact solution is offset by uncertainties in the optical properties of a real atmosphere, especially those of aerosols and the ground cover. Consequently, only the effects of single scattering are presented. The analysis that follows is simple, approximate, and empirical, yet quite accurate.

The proceeding formulation is based on the simple assumption that scattering by molecules and aerosols can be linearly separated. This is also called two-stream approximation.

Diffuse spectral irradiance on a horizontal surface, $D_{d\lambda}$ is composed of three parts, as follows:

$$D_{d\lambda} = DR + DA + DS \tag{22}$$

where DR is the diffuse spectral irradiance produced by Rayleigh scattering that arrives on the ground after the first pass through the atmosphere, DA is the diffuse spectral irradiance produced by aerosols that arrive on the ground after the first pass through the atmosphere, and DS is the diffuse spectral irradiance produced by multiple reflections.

The earth's atmosphere contains all the attenuators that have been treated so far and is considered a plane-parallel layer. If it is assumed that half the Rayleigh-scattered diffuse radiation generated by an irradiance of $I(0)\cos\theta$ on a horizontal surface at the top of the atmosphere is directed downwards, the diffuse irradiance due to Rayleigh scattering, D_R is as follows

$$D_R = I(0)\cos\theta\, \tau_{03}[1 - \tau_R]\tau_a/2. \tag{23}$$

It is appropriate at this moment to explain two phrases often employed by meteorologists : forward scatterance and backscatterance. The former implies an incident wave. This direction is usually toward the ground. Backscatterance is the fraction of the total scattered energy propagated in a direction opposite to the incident wave, and this direction is usually toward outer space. In the Rayleigh atmosphere treated above, it is assumed that the forward and backward scatterances are each equal to 0.5.

Aerosol scattering is mainly in the forward direction. Therefore, it is necessary to specify a ratio of forward to total energy scattered, that is, forward scatterance, β_a and single scattering albedo, w_0. The diffuse irradiance due to aerosol scattering reaching the ground after the first pass through the atmosphere, D_A is as follows

$$D_A = I(0)\cos\theta\,\tau_{03}\tau_R[\,1 - \tau_a\,]\beta_a w_0. \quad (24)$$

The factor β_a is a function of the zenith angle(Table 2[36]).

Diffuse radiation arriving on the ground after the first pass through the atmosphere and direct radiation are in part reflected by the ground. This upwelling radiation is partly reflected back to the ground by the atmosphere. This process continues ad infinitum. These multiple reflections between the ground and the atmosphere(not to be confused with multiple scattering) add to the diffuse radiation reaching the ground after the first pass through the atmosphere. Diffuse irradiance due to multiple reflections, Ds

$$Ds = \frac{\alpha\beta(I+D_R+D_A)}{1-\alpha\beta} \quad (25)$$

where α is the surface albedo(0.05: Bowker[37]), and β is the atmospheric backscatter for radiation reflected from the surface.

3.1.3. Spectral Integration Model for cloudless skies

UV-B irradiance on a horizontal surface at the mean sun-earth distance can be obtained by integrating Eq. (5).

$$G_0(\lambda) = \sum_{\lambda=280}^{\lambda=320} (I_0 + D_R + D_A + D_S)\wedge\lambda$$
$$= \sum_{\lambda=280}^{\lambda=320} (I(0)\cos\theta + D_R + D_A)/(1-\alpha\beta))\wedge\lambda \quad (26)$$

Table 2. Ratio of forward to total scattering βa by aerosol.

Solar zenith angle	0	26	37	46	53	60	66	73	79
β_a	0.92	0.91	0.89	0.86	0.83	0.78	0.71	0.67	0.6

3.2. Surface UV-B irradiance under cloudy skies

3.2.1. The Mac model

This model uses surface observations of cloud layer amounts and types and has been widely used in North America to estimate solar radiation(Davis and McKay, 1989). UV-B spectral irradiance $G(\lambda)$ in a cloudy sky is given by

$$G(\lambda) = G_0(\lambda)\Pi[(1-C_i) + t_iC_i] \quad (27)$$

where $G_0(\lambda)$ is a theoretical cloudless sky flux, C_i is the observed cloud layer amount or opacity in the ith layer, t_i is the cloud transmissivity for the layer cloud type, and the cloud field transmission for each cloud layer has a cloudless part $(1 - C_i)$ and a cloudy part t_iC_i. Cloud transmissivities were obtained from

$$t_i = a_i\exp(-b_im_r) \quad (28)$$

where m_r is the relative optical air mass, a_i and b_i are empirically determined parameters for different cloud types, as shown in Table 3.

Table 3. Cloud transmissivity parameters.[38]

Cloud type	a	b
Altocumulus	0.556	0.053
Altostratus	0.413	0.004
Cirrocumulus	0.923	0.089
Cirrostratus	0.923	0.089
Cirrus	0.871	0.020
Cumulonimbus	0.119	-0.226
Cumulus	0.368	0.045
Stratus	0.252	0.100
Nimbostratus	0.119	-0.226
Stratocumulus	0.368	0.045
Fog	0.163	-0.031

The albedo of the atmosphere for surface reflected radiation can be expressed as

$$\beta = \beta(R)(1-C) + \beta(a) + \beta_c C \quad (29)$$

$$\beta(a) = [1-\tau_a]w_0(1-\beta_a) \quad (30)$$

where $\beta(R)$ is the sum of the components due to molecular scattering assumed to apply only to the cloudless portion of the sky = 0.0685 (Lacis *et al.*, 1974), $\beta(a)$ is scattering by aerosol in the atmosphere below the cloud base, β_c is the average cloud albedo = 0.6 (Suckling and Hay, 1977), C is the total cloud amount, τ_a is the transmissions after absorption and scattering by aerosol, β_a is the ratio of forward to total scattering, and w_0 is the single scattering albedo.

3.2.2. Total cloud-based model

The total cloud -based model computed by this study under cloudy skies is expressed by

$$G(\lambda) = G_0(\lambda)[(1-C) + tC] \quad (31)$$

where t is the mean cloud transmissivity for the layer cloud type

3.3. Calculation of DOSE

The biological dose of UV-B at the surface is given by an integration of the spectral irradiance $G(\lambda)_0$ and an erythemal action spectrum $E(\lambda)$ given by McKinlay and Diffey[24)] shown in Fig. 2 for a particular biological effect over all wavelengths and a time period.

$$\text{UV DOSE} = \iint G(\lambda)E(\lambda)d\lambda\, dt \quad (32)$$

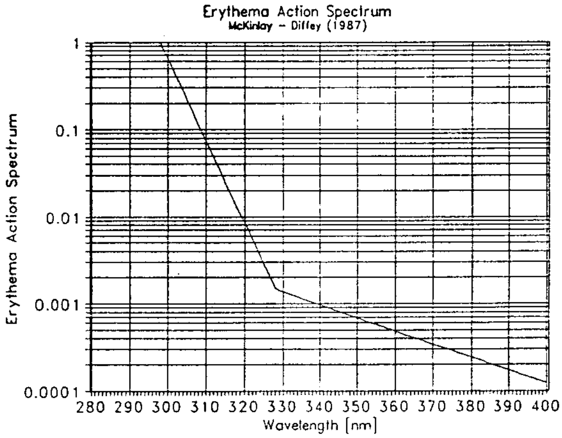

Fig. 2. Erythema action spectrum.

3.4. Error assessment

The model will contain errors due to inaccuracies in the model itself and in the input data. The performance of the model in estimating irradiances for periods of a day or longer will be discussed using the mean bias error(MBE) and root mean square error(RMSE). They are expressed as

$$\text{MBE} = \sum_{i=0}^{N}(P_i - M_i)/N \quad (33)$$

$$\text{RMSE} = [\sum_{i=0}^{N}(P_i - M_i)^2/N]^{1/2} \quad (34)$$

where P_i and M_i are the ith predicted and measured irradiance values.

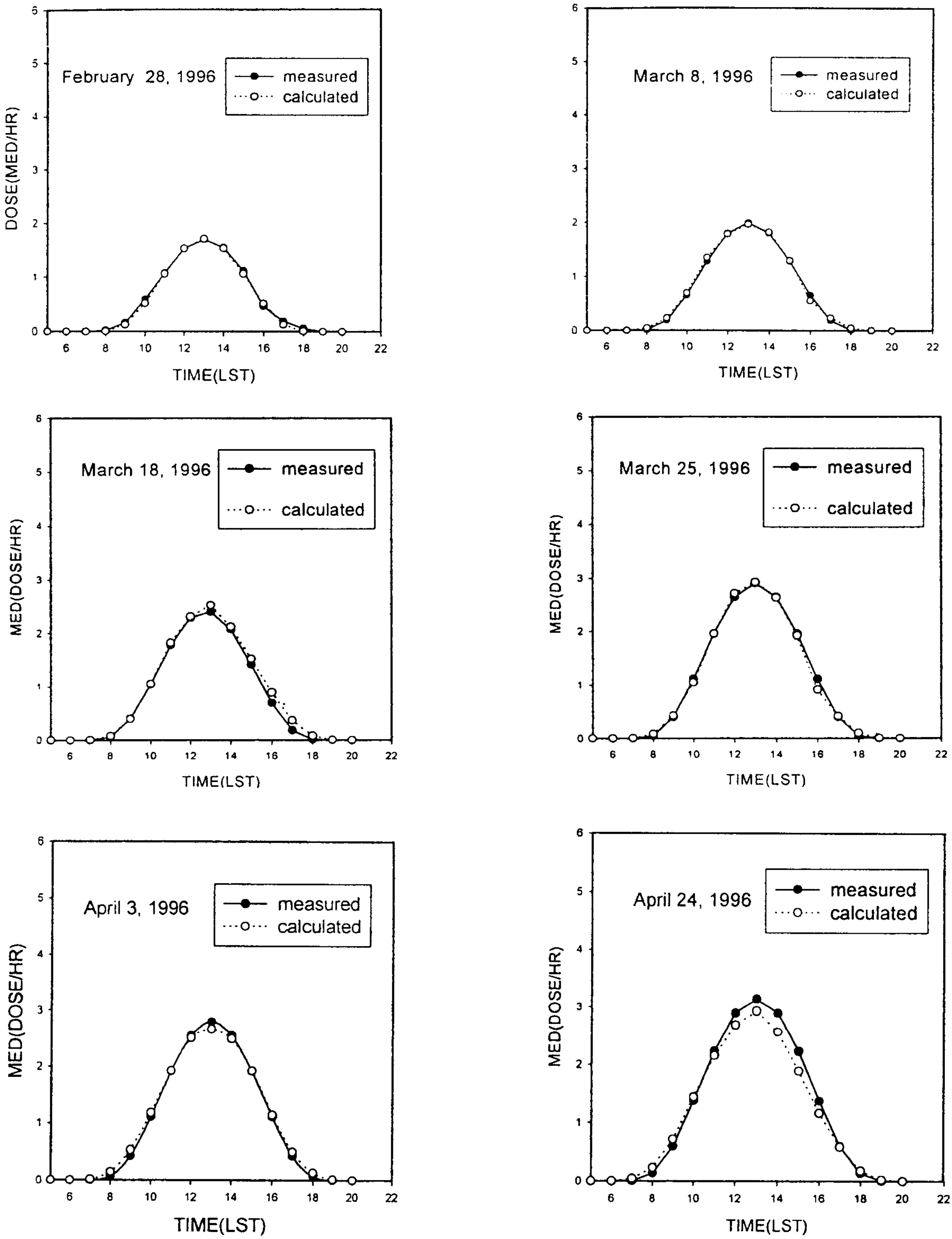

Fig. 3. Diurnal variation of calculated(dash and circle) and measured(line and circle) UV-B MED for some cloudless days.

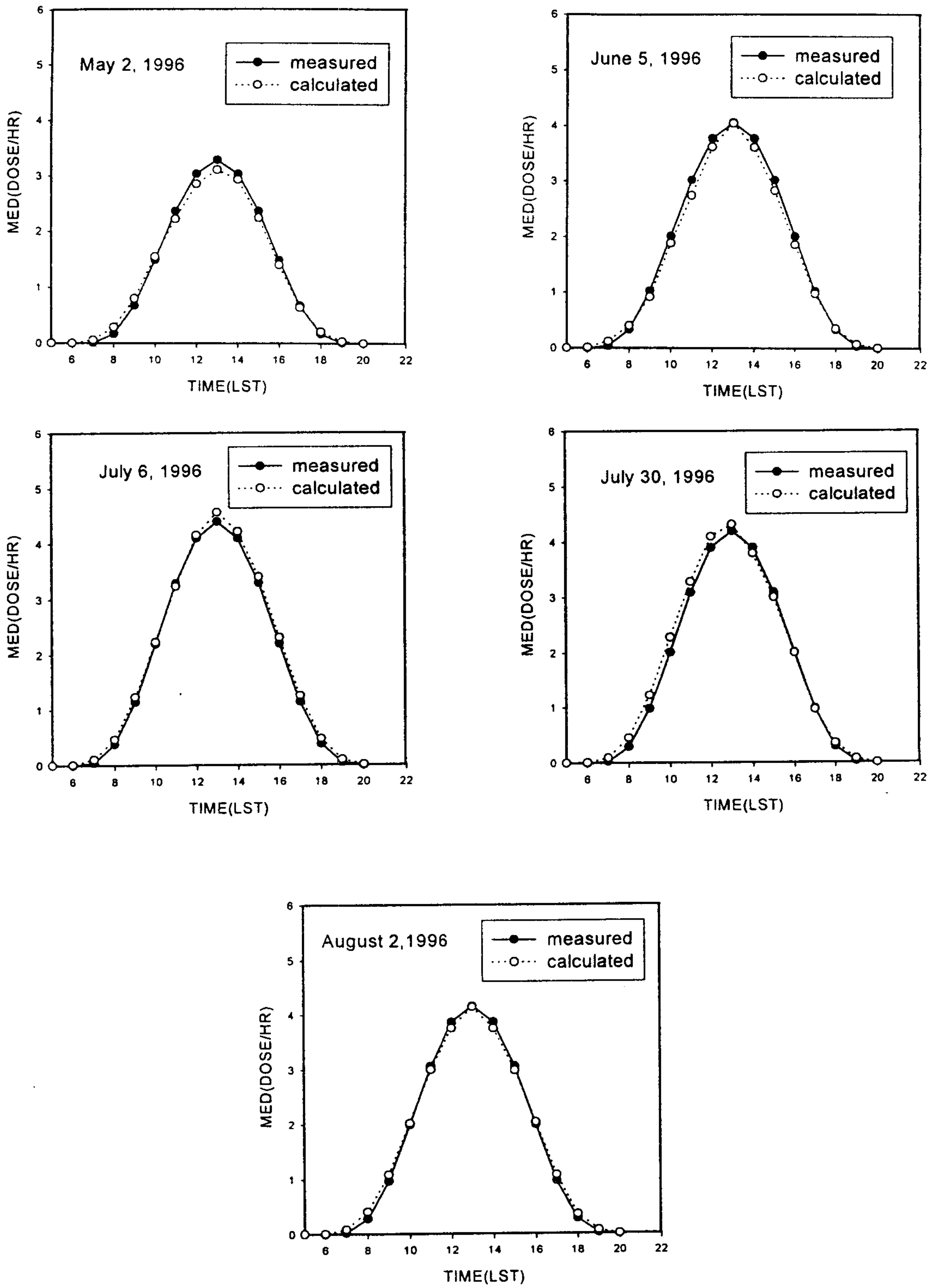

Fig. 3. (Continued.)

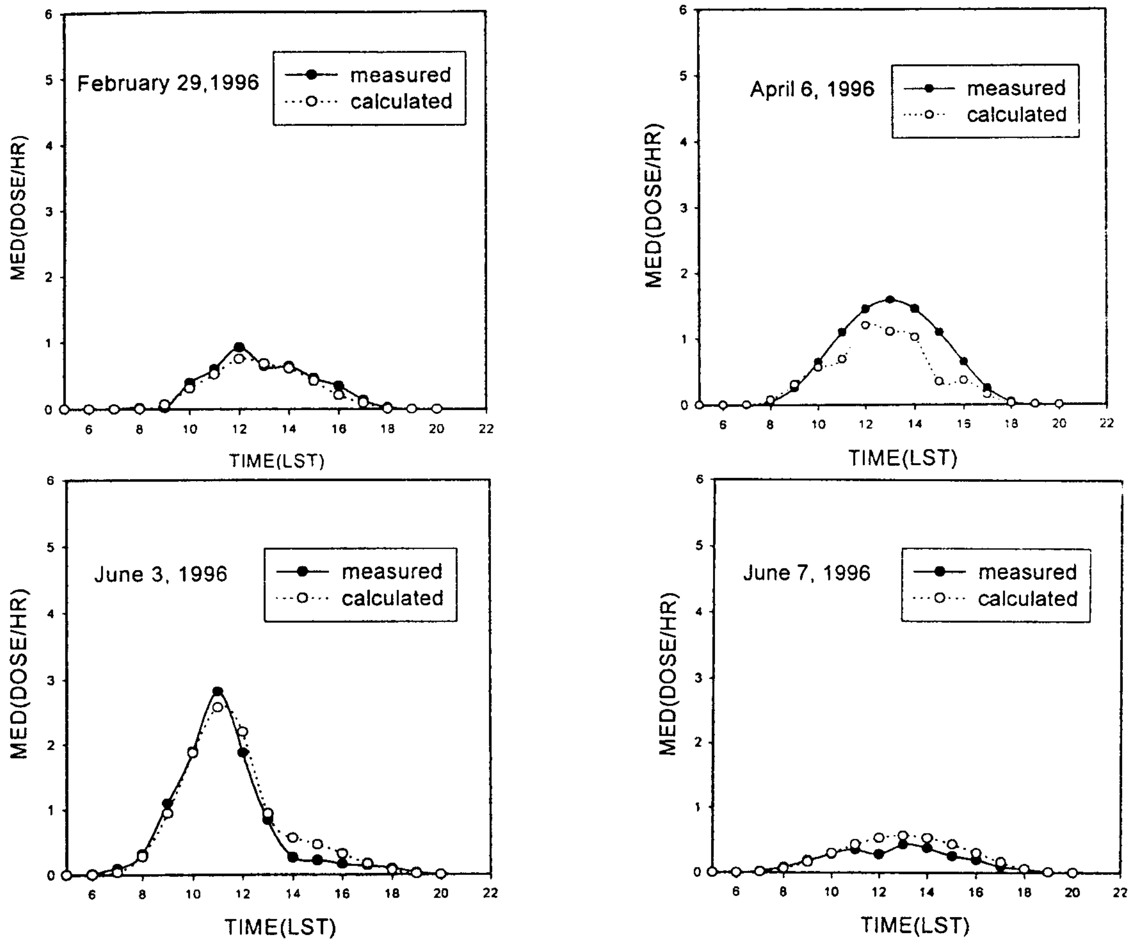

Fig. 4. Same as Fig. 3 except for some cloudy days.

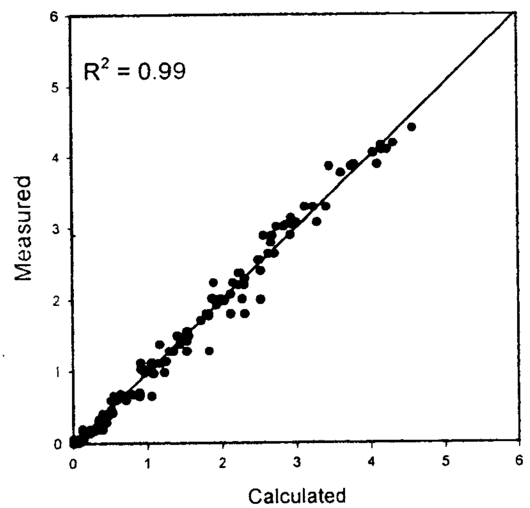

Fig. 5. Correlation between calculated and measured daily UV-B MED of cloudless skies.

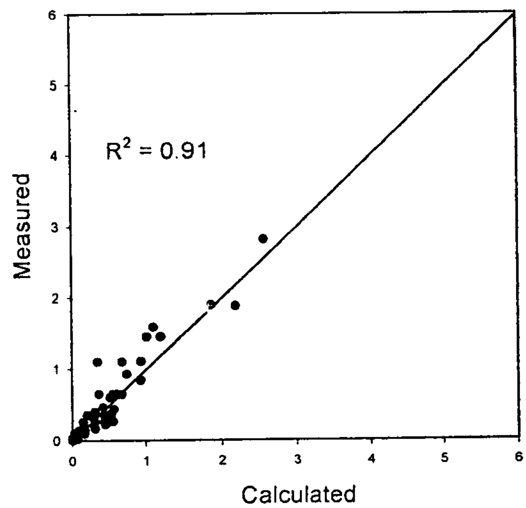

Fig. 6. Same as Fig. 5 except for some cloudy days.

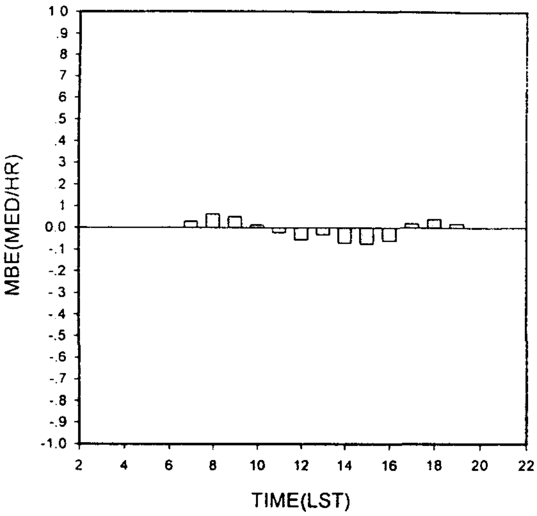

Fig. 7. Variation of MBE for cloudless sky UY-B MED.

4. Results

4.1. Cloudless skies

Cloudless skies were determined on the days when the interpolated total cloud amount for all measurement times was zero. Cloudless sky irradiances were calculated for spring and summer. The calculated and measured hourly irradiances for certain days are compared in Fig. 3.

A fixed value for Angstrom's wavelength exponent(αw = 1.3) and variable value for the single scattering albedo(ω_0 = 0.9) were assumed.

The results of this exercise are shown in Fig. 3, where visibilities and total ozone amounts were used as measured values. The calculations and measurements agreed within about a 1% error.
This difference is attributed to the effect of stray light on UV-Biometer measurements. The determined coefficient(R^2) value on the regression line related to the measured and calculated values was about 99%(Fig. 5). The mean bias error(Fig. 7) was within 5%(measured irradiance larger than calculated near 1400 LST) of the mean measured hourly irradiance. The root mean square error(RMSE) was within 10% for hourly irradiance(Fig. 9), The highest RMSE was recorded at 1200 LST. Yet the result of the model showed a good agreement for the hourly MED.

Sensitivity to visibility, the total ozone amount, single scattering albedo, solar zenith angle, and surface albedo, etc has been previously studied by Kim *et al.*[15])

4.2. Cloudy skies

The calculated and measured hourly irradiances for certain cloudy days in 1996 are compared in Fig. 4. The determined coefficient value on the regression line relating the two sets of values was about 91%(Fig. 7). The mean bias error(Fig. 9) was within 10% (calculated irradiance larger than measured) of the mean measured hourly irradiance. The RMSE was within 15% for hourly irradiance(Fig.10).

Including the effects of clouds in radiation models poses a number of problems, however,

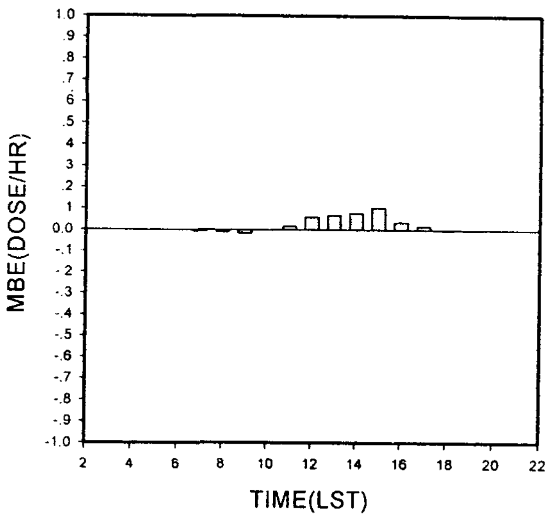

Fig. 8. Same as Fig. 7 except for cloudy sky.

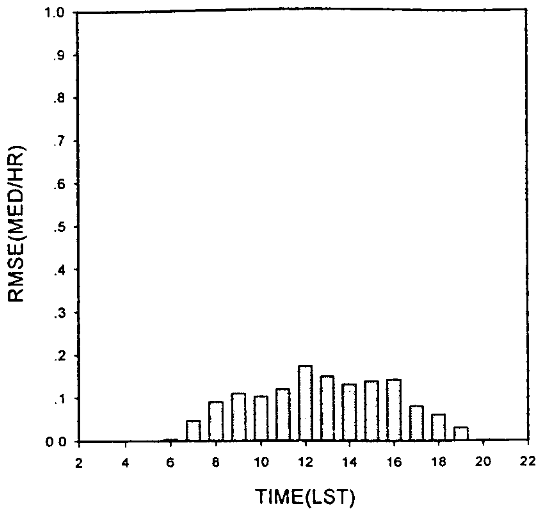

Fig. 9. Variation of RMSE for cloudless sky UV-B MED over averaging periods.

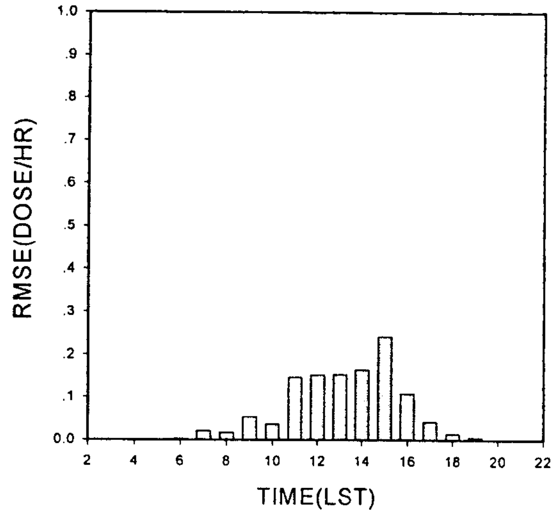

Fig. 10. Same as Fig. 9 except for cloudy sky.

they have been dealt with quite adquately.[25] Clouds reduce the accuracy of the computer model. The scarcity of observed data, the physical and dynamic nature of individual clouds and cloud fields, the variety of geometries and optical properties, and the occurrence of multiple layers all produce difficulties for modelling the effects of cloud on radiation.

There are usually three types of cloud observations; the amount of low clouds, the total amount of clouds, and the type of cloud at each layer. Since clouds are dynamic, observations are taken to represent only the half hour before and after the actual observation. Including observer and sampling errors, the measured irradiance under a cloud deck is assumed to have a low accuracy.[25]

Haurwitz's empirically-determined cloud transmissivities were determined for overcast conditions with single cloud types with no correction for overlying clouds which must have been present on some occasions. Therefore, the calculated transmissivity parameters for low and middle level clouds will most probably underestimate cloud transmissivity.

Accordingly, revised cloud transmissivities based on satellite measurements are need as well as regional aerosol climatologies-these are universal needs for applying any solar radiation model. Application is not seriously limited by the abundance of surface-based ozone measurements and incomplete and questionable surface cloud information since this information can be obtained from satellites.

5. Conclusion

The model discussed in this paper is one of the most useful modes tested in UV-B irradiance since it yields the best results overall and can be used at certain stations that make hourly cloud observations. Although Haurwitz's cloud transmission parameters have been used successfully in various parts of the world, they need to be validated with data from other sources. Consequently, this model was used in Pusan with observed cloud amounts for certain periods.

When the model was applied in Pusan it underestimated under cloudless skies and overestimated on cloudy days. The determined coefficient(R^2) values of cloudless and cloudy days on the regression line related to the measured and calculated values were about 99 % and 91 %, respectively. The hourly MBE values were between 5 and 10% for UV-B MED on an hourly basis. The hourly RMSE values were between 10 and 15%. Although errors in the calculated irradiance increased when cloud amounts were used, the results are still acceptable. It is important that the observed values of either the cloud layer amount or opacity are corrected using a satellite. This model can be easily extended to include action spectra, which are important in determining the possible effects of decreased ozone concentrations on life. The extended application of this model elsewhere is limited by the available ozone and cloud information. Furthermore, a numerical and statistical model to estimate solar radiation in the UV-B part should be developed.

Acknowledgement

We wish to thank Seong-Kyun Kim of the Korea Meteorological Administration. This study was supported by the Korea Science and Engineering Fundation(971-0404-014-2).

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