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[1] Department of Applied Physics, University of Athens, Athens, Greece
[2] Russian Academy of Sciences, St. Petersburg Research Center for Ecological Safety, St. Petersburg, Russia

An Estimation of the Surface Solar Ultraviolet Irradiance During an Extreme Total Ozone Minimum

M. Efstathiou[1], **C. Varotsos**[1], and **K. Ya. Kondratyev**[2]

With 2 Figures

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Summary

A simple theoretical algorithm has been employed to estimate the solar ultraviolet irradiance at Athens, Greece (38.7 °N, 23.4 °E) during, summertime 1993, a year of extreme total ozone minimum in the existing data record. This estimation has been performed by using total ozone measurements as derived by both ground-based and satellite instrumentation. The utilization of the present investigation will assist to the various assesments for the risk of human health from the biologically-effective doses of the solar ultraviolet radiation arrived at the earth's surface during that time period.

1. Introduction

There is no doubt that in the absence of other changes, reduction in stratospheric ozone will result in solar ultraviolet radiation (SUVR) increase, which is mainly due to its enhancement in the spectral region 290–320 nm (UV-B). However, the measurement of UV-B trend is challenging from the instrumental complexity and stability, so a conflict is observed at the results of long-term studies. Nevertheless at rural sites the observed increase in UV-B is comparable with that expected from total ozone change.

Madronich (1992) based on total ozone (TOZ) trend, suggested a theoretical method for the estimation of the enhancement of UV-B radiation, resulted from total ozone loss, over the globe. Analysis of spectral UV-B measurements at Lauder, New Zealand (45 °S) during 1990, in conjunction with total ozone measurements, showed that solar zenith angle (SZA) and cloud cover are among the dominant factors, which influence UV-B (McKenzie et al., 1991). Alongside it is well established in the literature that 1% TOZ decline results in an increase of the erythemally-active UV irradiance of 1.25± 0.20%. (WMO, 1994).

The purpose of the present note is to explore the amplification of UV-B irradiance induced from the extreme minimum of total ozone content observed in summertime period of 1993. It is of great importance due to the fact Athens is one of the most densely populated cities of the mid-latitudes, with high sunshine duration.

2. Data and Analysis

Recently, Varotsos et al., (1994) proposed a simple parametric model for the calculation of direct and diffuse SUVR reaching the earth's surface under different atmospheric conditions in dependence on time and geographic location. A brief description of this simple parametric model is given below.

2.1 Brief Description of Varotsos Model

2.1.1 Direct SUVR

According to Bird and Riordan (1986) the direct solar ultraviolet radiation reaching the earth's surface can be easily calculated by using the following expression:

$$I_d(\lambda) = S_o(\lambda)Ht_r(\lambda)t_a(\lambda)t_w(\lambda)t_o(\lambda)t_u(\lambda) \quad (1)$$

where, $I_d(\lambda)$ is the direct normal irradiance at the Earth's surface (at the wavelength λ), $S_o(\lambda)$ is the corresponding extraterrestrial solar irradiance at the mean Earth-Sun distance (Neckel and Labs, 1981). H is the correction factor for the change of the Earth-Sun distance throughout the year and it is given by the following equation:

$$H = 1.00011 + 0.034221\cos y + 0.001280\sin y + 0.000719\cos(2y) + 0.000077\sin(2y) \quad (2)$$

where y represents the day angle $[2\pi(d-1)/365, d$ is the number of days elapsed from the beginning of the year (Julian day)]. The transmission functions of the atmosphere at the wavelength λ for molecular (Rayleigh) scattering, aerosol attenuation (scattering and absorption), water vapor absorption, ozone absorption and uniformly mixed minor gases (CO_2, CO, CH_4, N_2O, O_2) absorption are denoted by the parameters $t_r(\lambda)$, $t_a(\lambda)$, $t_w(\lambda)$, $t_o(\lambda)$ and $t_u(\lambda)$, respectively. In the following, the transmission functions used for the calculation of the direct component of SUVR, are given:

- $$t_r(\lambda) = \exp\{-M'/[\lambda^4(115.6406 - 1.335/\lambda^2)]\} \quad (3)$$

Rayleigh scattering transmission function

- $$t_a(\lambda) = \exp(-b\lambda^{-\alpha}M) \quad (4)$$

Aerosol attenuation transmission function

- $$t_w(\lambda) = \exp[-0.2385a_w(\lambda)WM/(1 + 20.07a_w(\lambda)WM)^{0.45}] \quad (5)$$

Water vapor transmission function (Leckner, 1978)

- $$t_o(\lambda) = \exp[-a_o(\lambda)O_3M_o] \quad (6)$$

Ozone absorption transmission function (Leckner, 1978)

- $$t_u(\lambda) = \exp[-1.41a_u(\lambda)M'/(1 + 118.3a_u(\lambda)M')^{0.45}] \quad (7)$$

Uniformly mixed gases transmission function, where M' is the pressure-corrected relative optic air mass and M is the relative optic air mass given by the following expression:

$$M = [\cos z + 0.15(93.885 - z)^{-1.253}]^{-1} \quad (8)$$

In the latter expression z stands for the apparent solar zenith angle in rad and it is given by means of a computer algorithm for the calculation of the solar position with a precision to $1'$ of arc (Komhyr, 1980).

As far as the above mentioned parameter M', it is given by the equation:

$$M' = MP/P_o \quad (9)$$

where P is the measured surface pressure in hPa and $P_o = 1013$ hPa. In equation (4) α is the wavelength exponent and b is the Angstrom's turbidity coefficient that is given by the following expression:

$$b = [0.025 + 0.1\cos^2(\varphi)]\exp(-0.7h) \quad (10)$$

where φ and h is the latitude and altitude of the area of interest, respectively.

Apart from the above annual average of the Angstrom's turbidity coefficient versus the latitude, temporal variation of this variable can be accounted for by means of the experimental observations given in the following Table 1.

Table 1. *Temporal Variation of the Angstrom's Turbidity Coefficient Throughout the Year, Based Upon Experimental Observations*

Month	b
January	b / 1.22
February	b / 1.197
March	b / 1.0826
April	b / 0.89
May	b / 0.82
June	b / 0.71
July	b / 0.82
August	b / 0.89
September	b
October	b / 1.083
November	b / 1.14
December	b / 1.22

It should be noted that in this model the dependence of the Angstrom's turbidity coefficient has no taken into account on local conditions.

The wavelength exponent α in equation (4) takes the values 1.0274 and 1.206, when $\lambda < 0.5$ μm and $\lambda >= 0.5$ μm respectively.

In equations (5), (6) and (7), $a_w(\lambda)$ represents the water vapor absorption coefficient (at wavelength λ) and W the precipitable water vapor (in cm) in a vertical path. In the spectral region 0.28–0.75 μm, $a_w(\lambda) = 0$, so that $t_w(\lambda) = 1$. Consequently $a_o(\lambda)$ is the ozone absorption coefficient (at wavelength λ), O_3 is the total ozone amount in atm-cm and $a_u(\lambda)$ is a combination of an absorption coefficient and gaseous amount. Mo is the relative optical ozone mass which is given by the following equation (Bird and Riordan, 1986)

$$M_o = (1 + ho/6370)/[\cos^2 z + 2ho/6370]^{0.5} \tag{11}$$

where *ho* stands for the height of the maximum ozone concentration and it is calculated by means of following twelve expressions, which resulted from the examination of *ho* values for different latitudes and the whole year. It was identified a linear dependence of *ho* with latitude and it is presented at the following Table 2.

The equations of the Table 2 are valid for both hemispheres up to 70° latitude. In these equations φ stands for the latitude of the area of interest and it is given as an absolute value.

The values of $S_o(\lambda)$, $a_o(\lambda)$ and $a_u(\lambda)$ at various wavelengths are given by Bird and Riordan (1986).

Table 2. *Linear Dependence of ho with Latitude φ*

Month	*ho*
January	$ho = 35.919 + 0.0069$ (φ-30)
February	$ho = 36.043 + 0.0858$ (φ-30)
March	$ho = 33.799 + 0.1770$ (φ-30)
April	$ho = 34.048 + 0.6680$ (φ-30)
May	$ho = 31.610 + 0.1680$ (φ-30)
June	$ho = 31.667 + 0.1881$ (φ-30)
July	$ho = 34.238 + 0.1031$ (φ-30)
August	$ho = 34.186 + 0.0986$ (φ-30)
September	$ho = 34.069 + 0.1284$ (φ-30)
October	$ho = 33.909 + 0.1789$ (φ-30)
November	$ho = 33.734 + 0.2527$ (φ-30)
December	$ho = 35.959 + 0.0672$ (φ-30)

For the vertical ozone profile we have employed the model of Barnett and Corney (1985).

The direct solar irradiance at the Earth's surface on a horizontal plane is given by the multiplication of the direct normal irradiance at the Earth's surface $I_d(\lambda)$ by $\cos z$, where z is the apparent solar zenith angle.

Finally, to consider the percentage cloud amount in the sky, we use the following equation instead of the expression (1):

$$I_d'(\lambda) = [S_o(\lambda)Ht_r(\lambda)t_a(\lambda)t_w(\lambda)t_o(\lambda)t_u(\lambda)]C \tag{12}$$

For the factor c the following three different cases are specified (Diffey, 1982): $C = 1$, 0.7, 0.2 for clear sky, partially cloudy sky, overcast sky respectively.

2.1.2 Diffuse SUVR

The total diffuse irradiance on a horizontal surface at the ground is usually given by the equation

$$I_s(\lambda) = I_r(\lambda) + I_a(\lambda) + I_g(\lambda) \tag{13}$$

where, $I_s(\lambda)$ is the total diffuse irradiance on a horizontal surface at the ground, (at wavelength λ), $I_r(\lambda)$ is the Rayleigh scattering component, $I_a(\lambda)$ is the aerosol scattering component, and $I_g(\lambda)$ is the component taking into account the multiple reflected radiation between the air and the ground.

Justus and Paris (1985) proposed a model for the calculation of the diffuse SUVR, that was later modified by Bird and Riordan (1986). By means of this model we calculate the diffuse irradiance.

Instead of equation (12) we may use the following equation that takes into account the percentage cloud amount in the sky

$$I_s'(\lambda) = [I_r(\lambda) + I_a(\lambda) + I_g(\lambda)]C \tag{14}$$

In order to determine the multiple reflected radiation, it is given special care in the use of the appropriate ground albedo value. The following values are employed:

Table 3. *Values of Ground Albedo*

Fresh snow	0.85
Ice	0.35
Sand	0.25
Grassy field	0.2
Dry, ploughed field	0.125
Water	0.1
Forest	0.065

In the case of Athens we used the value 0.3 for mean ground albedo.

2.1.3 Total SUVR

Considering the aforementioned analysis for the calculation of the two SUVR components, the amount of the total solar irradiance on a horizontal surface at the ground yields:

$$I_t(\lambda) = I_d(\lambda)\cos z + I_s(\lambda) \tag{14}$$

where, $I_t(\lambda)$ is the total solar irradiance on a horizontal surface at the ground, at a wavelength λ and z is the apparent solar zenith angle.

In the case of Athens we applied the above described model by employing meteorological data collected at the Athens University Ozone Station.

2.2 Total Ozone Data

Daily column ozone observations (version7) made by TOMS instrument on board Nimbus-7

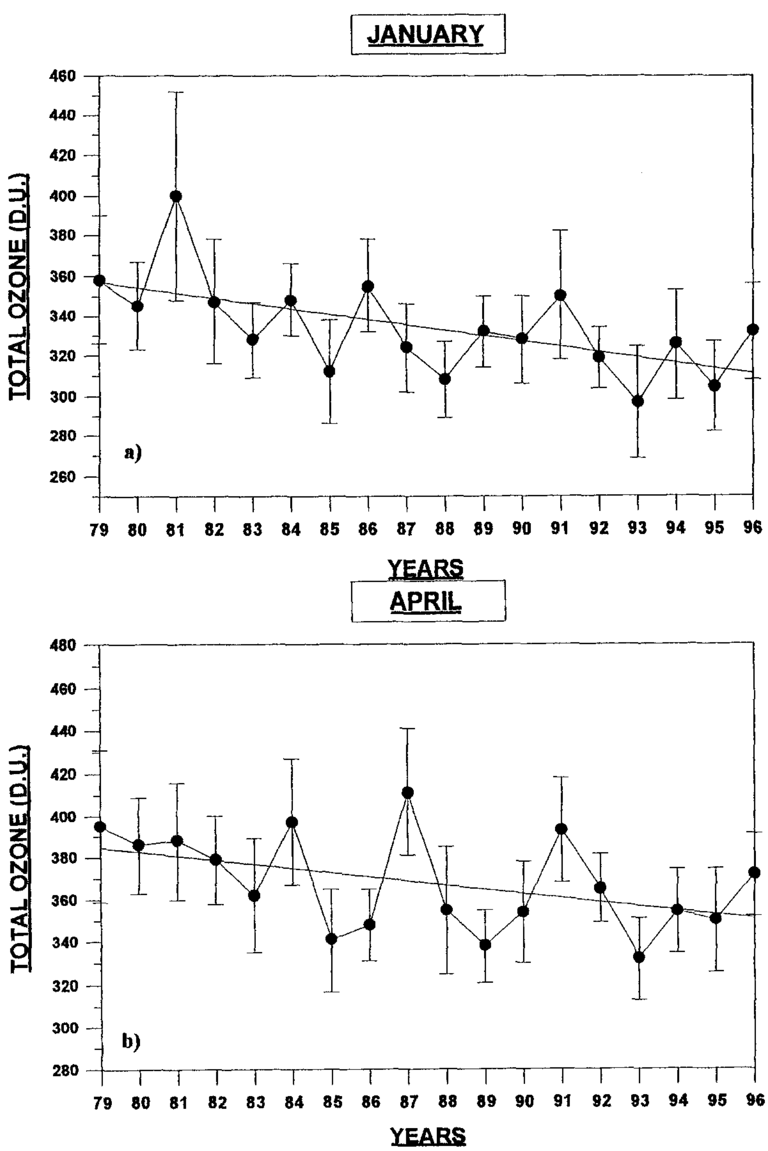

Fig. 1. Monthly mean and standard deviation of TOZ values deduced from TOMS and Dobson instrument at Athens (Greece) for January (a) and April (b)

satellite (Jan. 1979–April 1993) and Dobson#118 spectrophotometer at Athens, Greece, (Nov. 1991–Dec. 1996) were considered as input values in the model. It should be noted that the very good agreement of the TOZ observations made by the two instruments, during the common period of their operation justifies the consideration that the entire time series of 18 years of data is practically homogeneous.

3. Results and Discussion

3.1 Total Ozone Depletion over Athens

The results of a linear trend analysis, applied to monthly mean TOZ values deduced from daily measurements over Athens, for the mid-wintertime month (January) and the mid-springtime month (April) are shown in Fig. 1(a, b). It is evident from the Fig. 1 that the percentage total ozone depletion varies from 15% on January to 10% on April approximately. Recent studies Chandra et al. (1996) showed that the interannual variability has strong influence in the calculated TOZ depletion rates. Recently Chandra and Varotsos (1995), have shown that the dynamically induced component to the trend in TOZ accounts for 5% and 3.3%, for January and April, respectively. Considering this influence, the net ozone loss is significantly smaller than that shown in Fig. 1. An inspection of Fig. 1, denotes that an extreme total ozone minimum occurred during the transition period from winter to summer of 1993. This event had no any local character, but it had been observed on a global scale.

It is therefore of great importance to extract all the available information for the increased levels of UV-B that expected to reach the earth's surface at that time period.

3.2 Amplification of UV-B

The employment of daily total ozone values at the above mentioned parametric model allows us to detect the fluctuation of mean monthly UV-B irradiance for the summer months of the time period 1993–1996 to the long term mean (time period 1979–1996). The results are shown in Fig. 2. where UV-B fluctuations are plotted together with corresponding TOZ. In Fig. 2. the ratio of both TOZ and broad-band UV-B (290–320 nm) to the long-term mean is shown. It is evident from the latter figure that the fluctuations in TOZ of the order of 3.3±1.9% resulted to a corresponding increase to modeled broad-band UV-B of 3.6±1.4%. Although the real measurements of broad-band UV-B obtained at the Athens station is influenced by the local air polution, a very good agreement has been observed between the experimental and theoretically derived observations.

4. Conclusions

The 1993 total ozone content has been characterized by the lowest values ever observed above Athens, since 1978. This sudden total

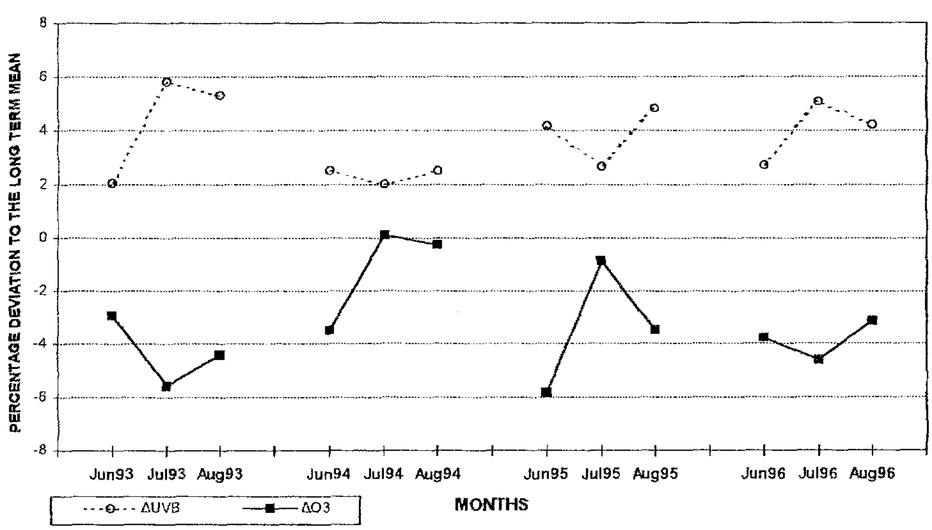

Fig. 2. The TOZ and UV-B at Athens, expressed as percentage deviation to the long-term mean, for summertime period (1993–1996)

ozone decline is consistent with the linear trend which has been defined in earlier studies. The theoretically deduced UV-B estimates are in close agreement with the real measurements. This fact supports the consideration that in general the severe ozone loss is accompanied with exceptional high UV-B doses.

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Author's addresses: M. Efstathiou and C. Varotsos, Department of Applied Physics, University of Athens, Panepistimioupolis, Build. PHYS-V, GR-15784, Greece; K. Ya. Kondratyev, Russian Academy of Sciences, St. Petersburg Research Center for Ecological Safety, Korpusnaya St., 18, 197110 St. Petersburg, Russia.