Aust. Met. Mag. 46 (1997) 179-184

Physics of stratospheric ozone and UV-B radiation

David J. Karoly
Cooperative Research Centre for Southern Hemisphere Meteorology
Monash University, Clayton, Australia

To help understand the role of ozone in protecting life on earth, a brief review is presented of the processes which determine the 'normal' distribution of ozone in the stratosphere and its variations with latitude, with elevation and with time of year. Next, the absorption of UV radiation by ozone is discussed, including the dependence of the surface exposure to UV radiation on solar elevation, ozone amount and cloud amount.

Introduction

Ozone in the stratosphere is vital for the health of all animal and plant life on earth because it absorbs most of the harmful ultraviolet (UV) radiation emitted by the sun before it reaches the earth's surface. To help understand the role of ozone in protecting life on earth, a brief review is presented of the processes which determine the 'normal' distribution of ozone in the stratosphere and its variations with latitude, with elevation and with time of year. Next, the absorption of UV radiation by ozone is discussed, including the dependence of the surface exposure to UV radiation on the angle of the sun's rays (solar elevation), ozone amount and cloud amount. This review is intended to set the scene for subsequent papers describing the chemistry of ozone and ozone depletion, the variability and trends in ozone on daily to interannual time-scales, and monitoring and prediction of surface exposure to UV radiation. There are a number of more comprehensive reviews of the distribution and trends in ozone and UV radiation, particularly the series of international assessments of ozone depletion (WMO 1986; 1989; 1992; 1995) and the earlier review for the Menzies Foundation (Plumb 1989). Interested readers are referred to those for further information.

Distribution of stratospheric ozone

Ozone (O_3, the triatomic form of the oxygen molecule) is very rare in the atmosphere but it plays vital roles in determining the thermal structure of the atmosphere and in protecting the earth's surface from the damaging effects of solar ultraviolet radiation. The absorption of UV radiation by ozone creates a heat source, which leads to a layer in the atmosphere in which temperature increases with height, called the stratosphere.

Ozone is found in two different height regions in the atmosphere. Most ozone is found in the stratosphere, at heights between about 10 and 50 km, in the 'ozone layer', while the remaining ozone is found closer to the surface, in the troposphere. This vertical distribution of ozone can be seen in Fig. 1, which shows the vertical profiles of ozone over Laverton (near Melbourne) and over Macquarie Island (in the Southern Ocean) on single days in September. The amount of ozone is commonly measured in three different ways:

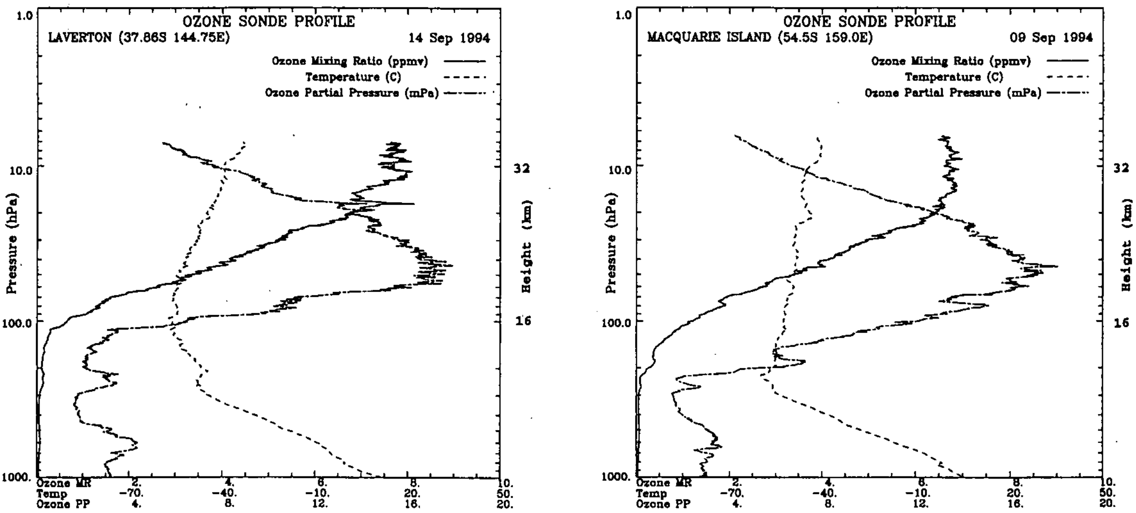
(a) as the ozone mixing ratio, in parts per million by volume (ppmv), which increases with height in the stratosphere up to about 35 km;

(b) as the ozone partial pressure, in mPa, which decreases with height above the lower stratosphere as the atmospheric pressure and density decreases; and

(c) as the total column ozone amount, in Dobson Units (DU, 300 DU = 3 mm of pure ozone at standard temperature and pressure, 0°C and 101 kPa).

Corresponding author address: Prof. D.J. Karoly, Cooperative Research Centre for Southern Hemisphere Meteorology, Monash University, Clayton, Vic. 3168, Australia.

Fig. 1 **Vertical profiles of ozone and temperature in the atmosphere from balloon-borne ozone sondes at (a) Laverton on 14 September 1994 and (b) Macquarie Island on 9 September 1994. Data are from the Atmosphere Watch Section, Bureau of Meteorology.**



The decreasing temperature with height in the troposphere and increasing temperature in the stratosphere are shown clearly in Fig. 1 also. The secondary maximum in the ozone partial pressure close to the earth's surface has some benefits due to its additional absorption of UV radiation but it has harmful effects on human health and on animal and plant tissue because ozone is toxic. Ground-level ozone concentrations are generally much smaller than those commonly found in the stratosphere.

There are two separate aspects to long-term changes in the ozone distribution; concerns about increases in tropospheric ozone and about decreases in stratospheric ozone, both of which are harmful to health. Ozone in the troposphere is important also as a greenhouse gas because it absorbs infrared radiation, so increasing tropospheric ozone may contribute to greenhouse climate change. In the remainder of this section, the processes determining the distribution of ozone in the stratosphere will be discussed, but the processes leading to long-term ozone depletion shall be left to a subsequent paper. There will be no further discussion of increases in tropospheric ozone. In any case, no significant increases in tropospheric ozone have been reported in the southern hemisphere (WMO 1995).

The main source of ozone in the atmosphere is photodissociation of molecular oxygen by high energy (short wavelength) UV radiation from the sun (Plumb 1989; Andrews et al. 1987):

$$\left.\begin{array}{r} O_2 + h\nu\ (\lambda < 242\ \mathrm{nm}) \rightarrow O + O \\ O + O_2 \rightarrow O_3 \end{array}\right\} \quad ...1$$

This process is limited by the amount of UV radiation because there is abundant molecular oxygen in the atmosphere. The production of ozone is faster over the equator, where there is more UV radiation, than at high latitudes, and at higher elevations, where less of the high-energy UV radiation has been absorbed by oxygen. However, the amount of ozone in any region depends on a balance between ozone loss processes and the transport of ozone by atmospheric motion, as well as ozone production. The variation of ozone as a function of height and latitude in Fig. 2 shows highest mixing ratios in the tropics in the summer hemisphere, with downward extension of this maximum into higher latitudes. There are rapid decreases in ozone above 40 km and below 20 km, associated with loss processes.

There are two dominant processes which destroy ozone in the atmosphere. The first is recombination of ozone with atomic oxygen (Plumb 1989; Andrews et al. 1987):

$$O_3 + O \rightarrow 2O_2 \quad ...2$$

The photochemical ozone production and loss cycle in the stratosphere represented by Eqns 1 and 2 is called the Chapman cycle. It is now recognised that this is much less efficient than catalytic cycles of the form:

$$\begin{array}{l} X + O_3 \rightarrow XO + O_2 \\ XO + O \rightarrow X + O_2 \end{array}$$

which may be summarised as

$$O + O_3 \overset{X}{\rightarrow} 2O_2 \quad ...3$$

Fig. 2 Distribution of mean ozone mixing ratio (in ppmv) in the stratosphere as a function of height and latitude in January and July. Values are obtained from five years of data from the Microwave Limb Sounder (MLS) instrument on the Upper Atmosphere Research Satellite (UARS) (Randel and Newman 1997).

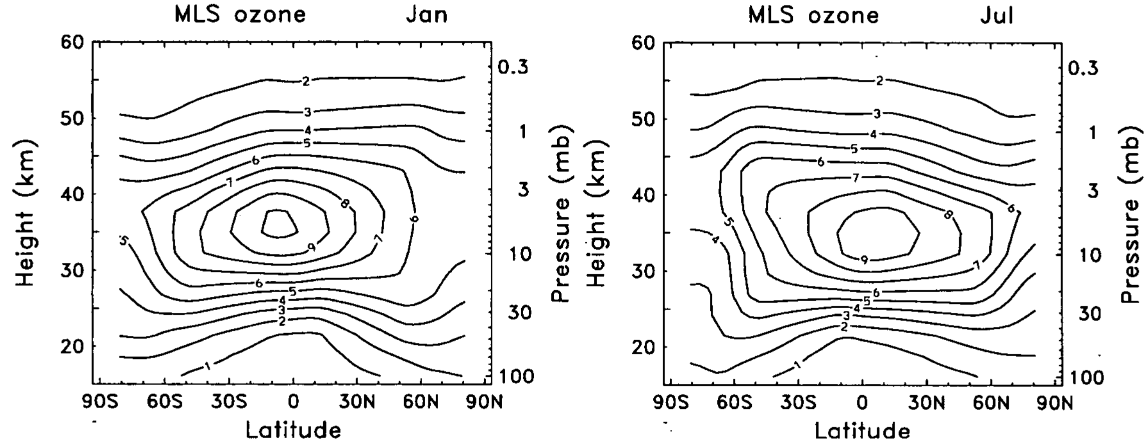
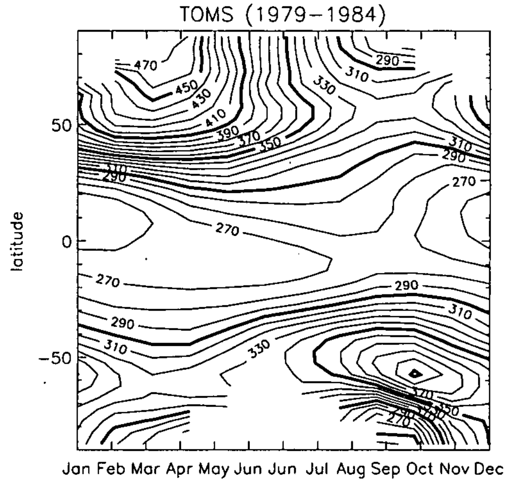


Fig. 3 Distribution of total column ozone amount (in Dobson Units) as a function of latitude and time of year. Values are averages over 1979-1984 of Total Ozone Mapping Spectrometer (TOMS) satellite observations (Randel and Newman 1997).



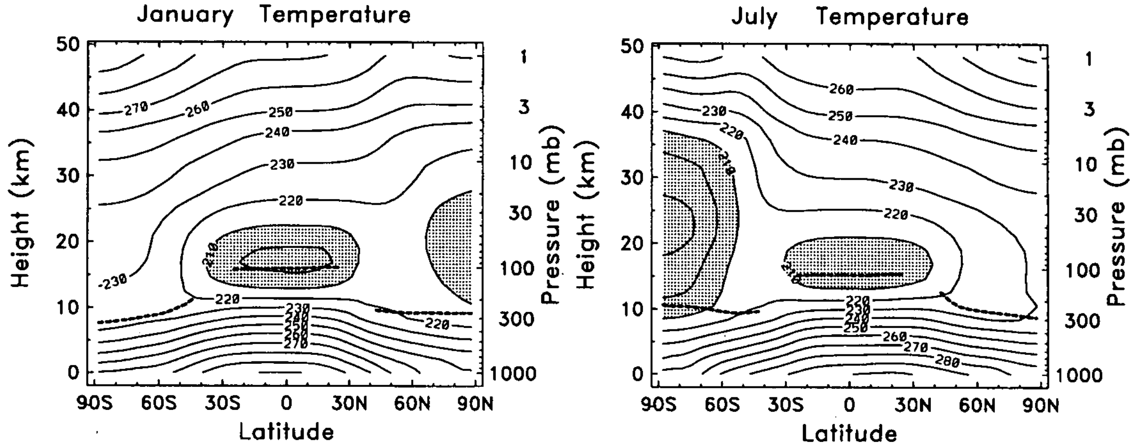
where the catalyst X may be one of the stratospheric radicals OH, H, NO, Cl, Br. This catalytic destruction of ozone in the stratosphere will be described further in the next paper on the chemistry of ozone and ozone depletion. Ozone is destroyed also in the troposphere through chemical reactions near the ground. The speed of this loss is determined by the rate of mixing of air from the stratosphere into the troposphere.

Since the density of air decreases rapidly with height, the maximum density of ozone is at lower levels than the maximum mixing ratio shown in Fig. 2. Below 25 km, mixing ratios are greater in middle and high latitudes than in the tropics. Hence, Fig. 2 is a misleading representation of the distribution of the total amount of ozone in a vertical column. The total column ozone amount shown in Fig. 3 is a minimum in the ozone source region in the tropics and a maximum at higher latitudes, particularly in early spring in each hemisphere. There is a relatively small seasonal cycle in the ozone amount in the tropics but a much larger seasonal cycle in middle and high latitudes, with the maximum ozone usually occurring in spring at most latitudes. This is a result of transport of ozone by atmospheric motion.

The dominant atmospheric circulation in the stratosphere is strong eastward motion (westerly winds) in the winter hemisphere and westward motion (easterly winds) in the summer hemisphere, as shown in Fig. 4. The mean zonal winds are in dynamical balance with the mean temperature structure, which is determined by radiative heating and cooling and by heat transport by the circulation (Andrews et al. 1987). The strongest westerly winds occur at high latitudes in the southern winter, in the region of strongest meridional (north-south) temperature gradient. The very cold temperatures at high latitudes in the southern winter stratosphere are an important factor in the existence of the Antarctic ozone hole in spring, which will be described in later papers. These strong zonal winds do not contribute to the poleward transport of ozone.

The poleward transport of ozone from its source in the tropics is provided by the mean poleward circulation

Fig. 4 Distribution of mean temperature and zonal wind in the atmosphere as functions of height and latitude in January and July. Values are obtained from long-term means of NMC (USA) analyses (Randel and Newman 1997).



in the stratosphere and by the net poleward transport in the large-scale wave disturbances which occur in the winter stratosphere (Plumb 1989; Andrews et al. 1987).

The effects of these two poleward transport processes are usually combined in the residual mean meridional circulation, shown schematically in Fig. 5. The mean upward motion in the tropics brings ozone-poor air from lower levels and transports ozone-rich air to higher latitudes. The downward and poleward transport in the winter hemisphere contributes to the ozone maximum there. There is also some mixing and diffusive transport by waves in the stratosphere which affects the ozone distribution.

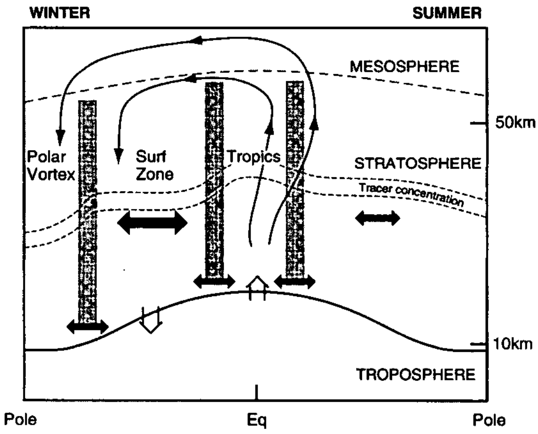
Hence, the ozone distribution is a balance between production, loss and transport. Changes to any of these processes can lead to net changes in the ozone distribution, such as the changes between summer and winter, or long-term trends. The observed trends in stratospheric ozone are described in the later paper by Atkinson (1997) and Fraser (1997) reviews the chemistry of ozone and ozone depletion.

UV radiation

In this section, a brief review is given of surface exposure to solar ultraviolet radiation and its dependence on solar elevation, ozone amount and other factors. A more comprehensive review is given in a later paper by Rikus (1997). Radiation is emitted by the sun over a range of wavelengths from the ultraviolet (UV, $\lambda < 400$ nm) to the infrared (IR, $\lambda > 700$ nm), with maximum intensity in the visible wavelengths (400 nm $< \lambda <$ 700 nm). Not all of this solar radiation which is incident on the top of the atmosphere reaches the surface, as much of it is absorbed or scattered in the atmosphere. High energy (short wavelength) UV radiation is dangerous to animal and plant life but, fortunately, most is absorbed in the atmosphere.

The UV wavelength range is further broken down into three subranges; UV-A 315-400 nm, UV-B 290-315 nm, and UV-C 220-290 nm, based on the different atmospheric absorption in each of these subranges (McKenzie et al. 1995). Some references use 320 nm as the separator between the UV-A and UV-B subranges. Essentially

Fig. 5 Schematic illustration of the residual mean meridional circulation (shown by the thin lines) and wave mixing (shown by the bold horizontal arrows) which determine meridional transport in the stratosphere.



all the radiation in the UV-C range is absorbed by molecular oxygen and ozone in the stratosphere in the photochemical reactions (Eqns 1 and 2) above. Radiation in the UV-B range is partly absorbed by ozone and is partly transmitted through the stratosphere, while radiation in the UV-A range is not absorbed in the stratosphere. Figure 6 shows the dependence of clear-sky UV radiation on wavelength for two different sites with different ozone amounts. Other trace gases in the stratosphere, such as sulphur dioxide emitted by volcanoes, can be important absorbers of UV radiation. This is discussed further by Rikus (1997).

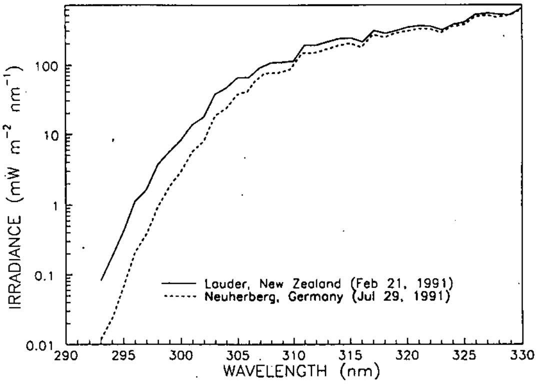
In the lower atmosphere, scattering of radiation by molecules, particles and water droplets is more important than absorption in determining how much UV-A and UV-B radiation reaches the surface. Light impinges on molecules or particles and is scattered into a range of angles, sending some back out to space. The effect of UV scattering is complex as it must account for the dependence of scattering on the type and size of the scatterer and allow for multiple scattering events. The amount of scattering increases with the number of particles in the air. The scattering of UV radiation means that a substantial fraction (up to 50 per cent) of the surface UV exposure comes from diffuse radiation (after scattering), in addition to direct rays. Hence, being shaded from direct sunlight provides only partial protection from UV exposure because of the diffuse (scattered) UV radiation.

Simple geometry influences the amount of UV radiation at the surface. The atmosphere acts as an effective filter through the absorption and scattering described above. The higher that the sun is in the sky (measured by the solar elevation), the shorter the atmospheric path for the radiation and less absorption and scattering can take place. Hence, UV radiation at the surface is most intense at local solar noon during mid-summer. Surface UV intensity depends on season, latitude and time of day because of this dependence on solar elevation.

Clouds are very important factors determining the amount of UV radiation, because the water droplets and ice particles in clouds are effective in scattering UV radiation. Cloud optical thickness is the primary factor determining UV radiative transfer through clouds and this is highly variable. Cloud optical thickness depends on the type, thickness, and amount of cloud and all can change rapidly with space and time. Since the effects of cloud on UV radiation are so variable and difficult to predict, it is more common to display clear-sky UV amounts, as in Fig. 6.

The sensitivity of different organisms to damage by exposure to UV radiation varies. This sensitivity is usually determined from laboratory experiments and represented as an action spectrum, the relative response to radiation at different wavelengths. For most plants and animals, the action spectrum shows much greater sensitivity to shorter wavelength UV radiation. The biologically effective UV radiation is determined by weighting the surface UV radiation by the biological action spectrum and integrating over all wavelengths. Multiplying this weighted surface UV exposure rate by 40 converts it from radiation amount in W m^{-2} to a standard UV index value, as recommended by the World Meteorological Organization (WMO). Several countries have developed operational forecasts of the UV index (Long et al. 1996; Wilson 1993) and an Australian system is described in a later paper in this volume (Rikus 1997). UV radiation exposure is considered to be minimal for UV index values less than two and very high for values higher than ten. Due to the dependence of UV radiation exposure on solar elevation, there is a large increase in the clear-sky UV index to a maximum each day at local solar noon. The mean monthly values of local noon clear-sky UV index over Australia are shown in Fig. 7 for December and July. These show much higher UV index values in summer than in winter, with very high index values being achieved over the whole country in summer. Even in winter, northern Australia still receives high doses of UV radiation.

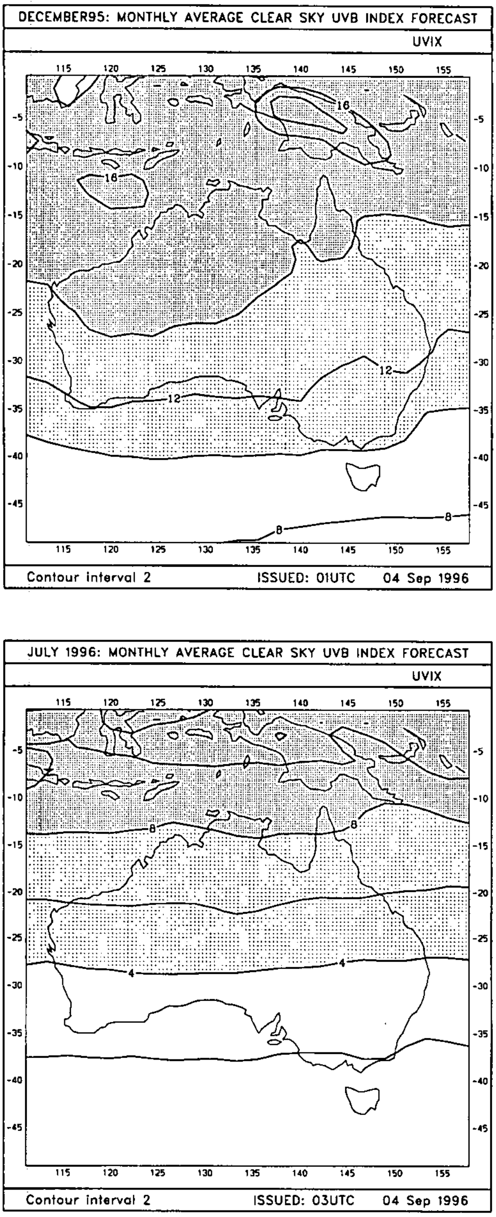
Fig. 6 **Measured clear-sky radiation (McKenzie et al. 1995) in New Zealand and Germany for solar zenith angle 34.3° . The ozone amount was 266 DU in New Zealand and 352 DU in Germany. A logarithmic scale is used for the irradiance on the y-axis.**



Conclusion

The distribution of ozone in the stratosphere and its role in absorbing most of the harmful UV radiation emitted by the sun has been reviewed briefly. The distribution of ozone in the stratosphere is the result of a delicate balance between photochemical production and loss, and transport by the atmospheric circulation. Surface expo-

Fig. 7 **Monthly mean local noon clear-sky UV index values over Australia for December 1995 and July 1996 obtained using a radiative transfer calculation and the daily ozone distribution, as described by Rikus (1997).**



sure to UV radiation depends on solar elevation, ozone amount and cloud amount. Even with no ozone depletion, there are high rates of exposure to UV radiation in Australia, particularly in northern Australia during winter (when there is little cloud), in central Australia at most times of the year, and in southern Australia during the summer half of the year. The health implications of this high UV radiation are important, even if there is no further ozone depletion. The recent observed ozone depletion exacerbates what is already a dangerous situation.

Acknowledgments

I would like to thank a number of colleagues for their assistance in preparing this review, including Roger Atkinson, Lilia Deschamps, Paul Fraser, Bill Randel, Lawrie Rikus and Darryn Waugh for providing the diagrams used and for their helpful comments.

References

Andrews, D. G., Holton, J.R. and Leovy, C.B. 1987. *Middle Atmosphere Dynamics*. Academic Press., 489pp.

Atkinson, R. J. 1997. Ozone variability over the Southern Hemisphere. *Aust. Met. Mag, 46*, 195-201.

Fraser, P. J. 1997. Chemistry of stratospheric ozone and ozone depletion. *Aust. Met. Mag, 46*, 185-193.

Long, C. S., Miller, A. J., Lee, H-T., Wild, J. D., Przywarty, R. C. and Hufford, D. 1996. Ultraviolet index forecasts issued by the National Weather Service. *Bull. Am. met. Soc., 77*, 729-48.

McKenzie, R. L., Blumthaler, M., Booth, C. R., Diaz, S. B., Frederick, J. E., Ito, T., Madronich, S. and Seckmeyer, G. 1995. Surface ultraviolet radiation. In Scientific Assessment of Ozone Depletion: 1994, *WMO Global Ozone Research and Monitoring Project Report No. 37*, 9.1-9.22.

Plumb, R. A. 1989. Atmospheric ozone - physics and chemistry. *Trans. Menzies Foundation, 15*, 3- 13.

Randel, W. J. and Newman, P. 1997 The stratosphere in the Southern Hemisphere. In *Meteorology of the Southern Hemisphere*, Amer. Meteor. Soc., in press.

Rikus, L. 1997. Prediction of UV-B dosage at the surface. *Aust. Met. Mag, 46*, 211-222.

Wilson, L. J. 1993. *Canada's UV Index – How It Is Computed and Disseminated.* Env. Canada, Atmos. Env. Service, 3pp.

WMO 1986. Atmospheric Ozone 1985, 3 Volumes, *WMO Global Ozone Research and Monitoring Project Report No. 16*, Geneva.

WMO 1989. Scientific Assessment of Stratospheric Ozone: 1989, 2 Volumes, *WMO Global Ozone Research and Monitoring Project Report No. 20*, Geneva.

WMO 1992. Scientific Assessment of Ozone Depletion: 1991, *WMO Global Ozone Research and Monitoring Project Report No. 25*, Geneva.

WMO 1995. Scientific Assessment of Ozone Depletion: 1994, *WMO Global Ozone Research and Monitoring Project Report No. 37*, Geneva.