UV climatology at McMurdo Station, Antarctica, based on version 2 data of the National Science Foundation’s Ultraviolet Radiation Monitoring Network

G. Bernhard, C. R. Booth, J. C. Ehramjian, and S. E. Nichol

Received 8 February 2005; revised 20 January 2006; accepted 15 February 2006; published 2 June 2006.

Spectral ultraviolet (UV) and visible irradiance has been measured near McMurdo Station, Antarctica, between 1989 and 2004 with a SUV-100 spectroradiometer. The instrument is part of the U.S. National Science Foundation’s UV Monitoring Network. Here we present a UV climatology for McMurdo based on the recently produced “version 2” data edition. Compared to the previously published “version 0” data set, version 2 data differ by −5 to 12% in the UV, depending on wavelength, solar zenith angle (SZA), and year. A comparison with results of a radiative transfer model confirmed that measurements of different years are consistent to within ±5%. Clear-sky spectra measured between October 1991 and March 1992 were significantly lower than spectra of other years because of the presence of volcanic aerosols. Total ozone column was calculated from UV spectra and was found in excellent agreement with collocated measurements of a Dobson spectrophotometer and satellite observations. Effective surface albedo was also estimated from clear-sky spectra. Monthly average albedo ranges between 0.69 for March and 0.84 for October. Biologically effective UV radiation is largest in November and December when low total ozone amounts coincide with relatively small SZAs. During these months, the noon-time UV Index typically varies between 2 and 5.5, but indices as high as 7.5 have been observed. The largest erythemal daily dose of 6.7 kJ/m² was measured on 28 November 1998. Linear regression analyses did not indicate statistically significant trends in UV nor visible radiation for the months September to January. For February and March, we found large, statistically significant positive trends in the UV and visible as well as for short-wave (0.3–3.0 μm) irradiance, ranging between 12 and 30% per decade. These trends are likely caused by changes in cloudiness and/or surface albedo, but the data do not allow unambiguous attribution of the increase to one of the two factors. On average, clouds reduce UV irradiance at 345 nm by 10% compared to clear-sky levels. Reductions vary substantially by month and year, can exceed 60% on rare occasions, and generally increase with wavelength. Between September and November, the variability in UV introduced by changes in total ozone is about twice as high as the UV variability due to clouds.


1. Introduction

[2] The National Science Foundation’s Office of Polar Programs (NSF/OPP) UV Monitoring Network was established in 1987 in response to severe ozone depletion reported in Antarctica. Biospherical Instruments (BSI) installed the first instruments in 1988 and has operated the network since [Booth et al., 1994]. The network currently consists of seven sites, which are mostly located at high latitudes. All stations measure global spectral irradiance between 280 and 600 nm with high-resolution SUV-100 spectroradiometers. Further information on the network and its data is available at the Website http://www.biospherical.com/NSF.

[3] Currently published network data are based on the originally released “version 0” data set. This data set has not been corrected for deviations of the angular response of SUV-100 spectroradiometers from the ideal cosine response nor for wavelength errors of approximately 0.1 nm, which affected spectra measured before 1997. To improve the accuracy and homogeneity of network data, a new data edition named “version 2” is currently being generated.
Version 2 data are not “pure” measurements like version 0 data, as radiative transfer calculations are used for the various corrections. These model calculations also provide clear-sky reference spectra during cloudy periods. Version 2 data from the instrument located at the South Pole (SPO) were introduced by Bernhard et al. [2004, hereinafter referred to as B04]. Here we present the new version 2 data set for the network instrument located near the U.S. Antarctic base McMurdo Station (MCM).

The UV climate at MCM differs in several important ways from that at SPO: at MCM, there is a pronounced diurnal variation in SZA (e.g., the SZA on 21 December varies between 54.4° and 78.7°, whereas the SZA at SPO is virtually constant at 66.6°); MCM has an annual cycle of surface albedo, whereas the albedo at SPO is constantly larger than 0.95; MCM is located at the coast and is surrounded by mountains, whereas SPO is located at 2830 m elevation on an effectively featureless plateau; clouds at McMurdo are more frequent than at SPO and normally have a larger optical depth. Background aerosol optical depths at the Antarctic coast tend to be higher than on the polar plateau [Shaw, 1982], and day-to-day variations in total column ozone during the annually recurring “ozone hole” are typically larger than at SPO.

2. Data Analysis

2.1. Measurements

Measurements of global spectral irradiance were performed at Arrival Heights (77°50′S, 166°40′E, 183 m above sea level) between December 1989 and January 2004 with a high-resolution SUV-100 spectroradiometer from BSI. The instrument is built into the roof of a modular building and measures solar spectra between 280 and 600 nm with a resolution of approximately 1.0 nm at a rate of 4 spectra per hour (1 spectrum per hour before 1997). Arrival Heights refers to a cliff-like location approximately 3 km north of McMurdo Station on the southern tip of Ross Island, which is surrounded by the Ross Sea and the Ross Ice Shelf. The active volcano Mount Erebus, 3795 m high, is 34 km north of the instrument. We use McMurdo rather than Arrival Heights when referring to instrument location and data because of the more familiar name recognition. The surface in the immediate vicinity of the instrument is covered with dark volcanic rocks, which are typically covered by snow. However, periods when the rock surface is exposed within an approximately 1 km radius around the instrument frequently occur between January and March. The area of the town of McMurdo is often free of snow also. According to Degünther et al. [1998], changes in surface albedo as far as 40 km away from the site of observation can significantly affect surface UV irradiance.

Version 2 data are based on “version 0 composite scans” that have been published by BSI on CD-ROM and/or the project’s website. The calibration of version 0 data is traceable to standards of spectral irradiance issued by the U.S. National Institute for Standards and Technology (NIST). The SUV-100 spectroradiometer is calibrated every two weeks. Detailed information on calibration procedures and quality control of version 0 data are provided in Operations Reports [Bernhard et al., 2003a]. The version 2 data set for MCM was processed in a similar way as the SPO data set, following the procedures outlined by B04 with some modifications to account for site specific differences explained below. Additional information is provided at the version 2 website (available at http://www.biospherical.com/nsf/Version2).

2.2. Radiative Transfer Calculations

Measurements were complemented with model calculations performed with the radiative transfer model UVSPEC/libRadtran version 1.0-beta, available at http://www.libradtran.org [Mayer et al., 1997; Mayer and Kylving, 2005]. The model’s pseudospherical disort radiative transfer solver with six streams was used.

The extraterrestrial spectrum (ETS) used for the modeling is based on the ETS recently proposed by Gueymard [2004]. Since the resolution of this spectrum is not sufficient to be used directly in the model, we superimposed the fine structure from the “Kitt Peak solar flux atlas” [Kurucz et al., 1984], measured with a Fourier Transform Spectroradiometer. The composite spectrum is identical with the spectrum $E_{\text{Gueymard}}(\lambda)$ introduced by B04.

Air density and pressure profiles were taken from Anderson et al. [1986]. Surface pressure at Arrival Heights was calculated from pressure data measured at McMurdo. Total column ozone was calculated from measured UV spectra as described in section 2.3. Ozone and temperature profiles were derived from ozone sonde observations performed at MCM [Kröger et al., 2003, and references therein]. These profiles are available on the database of the Network for the Detection of Stratospheric Change (http://www.ndsc.ws). Only profiles with a burst altitude of at least 30 km were used, and profiles were extrapolated to higher altitudes using an algorithm described by Bernhard et al. [2002]. On average, balloons were launched approximately every 2–3 days between mid-August and end of October in all years, and profiles closest in time with UV measurements were used for modeling. Only a few balloon measurements were available for the months November to April. We calculated the total ozone column from these profiles and selected those profiles for modeling whose total ozone amounts best matched the total ozone column observed by a Dobson spectrophotometer and satellite on the day of UV measurements.

Effective surface albedo was calculated from UV spectra measured during clear skies as described in section 2.6 and interpolated to times of cloudy days. Aerosol extinction was parameterized with Ångström’s turbidity formula by setting the Ångström coefficients $\alpha$ and $\beta$ to 1.5 and 0.00884, respectively. The resulting aerosol optical depth (AOD) at 500 nm is 0.025. This value is in agreement with measurements by Shaw [1982], who determined background AOD at McMurdo Station to $\tau = 0.025 \pm 0.010$, and measurements by Herber et al. [1993], which were performed at the coastal Antarctic stations Mirny (66°33′S, 93°00′E), Georg Forster (70°46′S, 11°41′E), and Molodeznaya (67°41′S, 45°54′E). The values of the Ångström coefficient $\alpha$ is based on measurements by NOAA’s Climate Monitoring and Diagnostics Laboratory (CMDL) performed at the SPO with the wavelength pair 412/675 nm [Climate Monitoring and Diagnostics Laboratory (CMDL), 2004]. Aerosol optical depth used in model calculations are too small for the period of September 1991 and 1994, when
volcanic aerosols from the eruptions of Cerro Hudson and Mt. Pinatubo led to substantial increases in aerosol optical depth. The effect of volcanic aerosols on UV is discussed in section 3.

Model spectra were convolved with a triangular function of 1.0 nm full width at half maximum (FWHM). Clear-sky spectra were selected from the data set on the basis of temporal variability of spectral irradiance at 600 nm using a similar method as described by B04.

2.3. Calculation of Total Column Ozone and Comparison With Dobson and TOMS Observations

Total column ozone was calculated from measured UV spectra using the inversion algorithm described by Bernhard et al. [2003b], the Bass and Paur [1985] ozone absorption cross section, and ozone and temperature profiles introduced in section 2.2. The ozone value returned by the algorithm was used for modeling. Figure 1 shows a comparison of these total ozone values with (1) measurements of a Dobson spectrophotometer operated at Arrival Heights by New Zealand’s National Institute of Water and Atmospheric Research (NIWA) and (2) version 8 overpass ozone data from NASA's Total Ozone Mapping Spectrometers (TOMS) installed on Nimbus-7 and Earth Probe satellites. Further statistics of the comparison can be found in Table 1. SUV-100 and Dobson total ozone values agree on average to within 1.000 ± 0.040 (±1σ). The comparison virtually does not depend on SZA and total ozone. Nimbus-7 TOMS and SUV-100 data show a similarly good agreement. Earth Probe TOMS data tend to be low by about 3%. Some of this difference may be related to the recent degradation of the Earth Probe TOMS front scan mirror (personal communication with R.D. McPeters, NASA). Both Nimbus-7 and Earth Probe TOMS data tend to be lower than SUV-100 data when total ozone is low. We decided to use our ozone data for modeling as they are available for every day with spectral measurements. The effect of total ozone on the model-based corrections discussed in the next section is small. For example, changing total ozone in the model by 10% typically changes corrected UV spectra by less than 0.1%.

2.4. Corrections Applied to Version 2 Data

Wavelength shifts have been corrected by correlating the Fraunhofer line structure in measured spectra to the corresponding structure in modeled spectra as described by B04. Residual wavelength uncertainties after correction are on average ±0.04 nm (±1σ) at 300 nm, and ±0.03 nm at longer wavelengths.

Corrections for the cosine error of the SUV-100 spectroradiometer have been performed with the method described by B04. These corrections pose a challenge since characterizations of the error prior to deployment of the SUV-100 proved to be not applicable to the assembled and installed instrument. Solar data further suggest that the error depended on azimuth angle and wavelength prior to a modification of the instrument’s irradiance collector in 2000. The error has also changed when the instrument was serviced. This results in relatively large uncertainties of the correction, particularly at large SZA.

The bandwidth of SUV-100 spectroradiometers varies between 0.9 and 1.1 nm in the UV-B and 0.7 and 0.9 nm in the visible. Processing of version 2 data involved normalization of all spectra to a uniform bandwidth of 1.0 nm and resampling of spectra to a uniform wavelength grid, see B04.

2.5. Uncertainty Budget

The total uncertainty of version 2 MCM data is composed of uncertainties related to (1) radiometric calibration and stability, (2) cosine and azimuthal error correction, (3) the finite resolution of the spectroradiometer, (4) residual wavelength errors after wavelength correction, (5) nonlinearity, (6) stray light, and (7) photon and electronic noise. All error sources have been discussed in detail by B04.

Table 1. Statistics of Comparison of SUV-100, Dobson, and TOMS Total Ozone Data at McMurdo

<table>
<thead>
<tr>
<th>Data Set</th>
<th>Ratio ± 1σ</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>NIWA Dobson/SUV-100</strong></td>
<td></td>
</tr>
<tr>
<td>All data</td>
<td>1.000 ± 0.040</td>
</tr>
<tr>
<td>SZA &lt; 80°</td>
<td>1.001 ± 0.037</td>
</tr>
<tr>
<td>Dobson AD pair</td>
<td>1.003 ± 0.033</td>
</tr>
<tr>
<td>Dobson CD pair</td>
<td>0.992 ± 0.055</td>
</tr>
<tr>
<td>Ozone &lt; 220 DU</td>
<td>0.986 ± 0.046</td>
</tr>
<tr>
<td>Ozone &gt; 220 DU</td>
<td>1.006 ± 0.036</td>
</tr>
<tr>
<td><strong>Nimbus-7 TOMS Version 8/SUV-100</strong> (December 1989 to April 1993)</td>
<td></td>
</tr>
<tr>
<td>All data</td>
<td>1.008 ± 0.051</td>
</tr>
<tr>
<td>SZA &lt; 80°</td>
<td>1.004 ± 0.041</td>
</tr>
<tr>
<td>Ozone &lt; 220 DU</td>
<td>0.967 ± 0.042</td>
</tr>
<tr>
<td>Ozone &gt; 220 DU</td>
<td>1.016 ± 0.049</td>
</tr>
<tr>
<td><strong>Earth Probe TOMS Version 8/SUV-100</strong> (August 1996 to January 2004)</td>
<td></td>
</tr>
<tr>
<td>All data</td>
<td>0.973 ± 0.055</td>
</tr>
<tr>
<td>SZA &lt; 80°</td>
<td>0.969 ± 0.044</td>
</tr>
<tr>
<td>Ozone &lt; 220 DU</td>
<td>0.944 ± 0.055</td>
</tr>
<tr>
<td>Ozone &gt; 220 DU</td>
<td>0.984 ± 0.051</td>
</tr>
</tbody>
</table>
for the instrument at SPO. Because the SUV-100 radiometers installed at MCM and SPO are virtually identical, we adopted the uncertainties of errors sources 1, 3, 5, 6 and 7 reported by B04 without modification for the instrument at MCM. Uncertainties related to source 4 were recalculated and are slightly different from those of the SPO instrument. The detection limit or “noise equivalent irradiance” (NEI) was also recalculated and varied between 0.00039 W/(cm² nm) and 0.00113 W/(cm² nm); the average NEI was 0.00077 W/(cm² nm). Uncertainties related to source 2 are site specific, and were recalculated on the basis of prevailing atmospheric conditions, surface albedo, and SZA at MCM. Detailed information on uncertainty components contributing to error source 2 at MCM is provided on the version 2 website (available at http://www.biospherical.com/nsf/Version2). Uncertainties due to source 2 are somewhat smaller for data of volumes 1–9 (1989–1999) than for data of volumes 10–13 (2000–2004), particularly at long wavelengths. This is due to the fact that azimuthal errors (i.e., variations of system sensitivity with the azimuth position of the Sun), which mostly affected data of earlier years, can be corrected with less uncertainty than cosine errors, which increased after the collector upgrade in 2000.

The combined uncertainty of version 2 MCM data was calculated and is presented in Table 2. Expanded relative uncertainties (coverage factor \(k = 2\)) vary between 4.6% and 15.8%. Expanded uncertainties for spectral irradiance at 310 nm range between 6.4% and 6.8% and are dominated by uncertainties related to calibration, stability, and wavelength errors. At larger wavelengths, the greatest uncertainty arises from the insufficient knowledge of the contribution of direct irradiance to global irradiance. This is a problem particularly during periods of varying cloudiness when it is difficult to determine whether the cosine correction factor for clear or overcast sky should be applied. The uncertainty budget presented in Table 2 therefore distinguishes between the cases of “clear-sky,” “overcast” and “unknown sky condition.” Expanded uncertainties for spectral irradiance at 600 nm range between 4.6% for overcast conditions and 15.8% for periods with variable cloudiness and large SZA. Expanded uncertainties for erythemal irradiance (i.e., spectral irradiance weighted with the CIE action spectrum for sunburn [McKinlay and Diffey, 1993]) is also provided in Table 2.

Table 2. Uncertainty Budget of Version 2 Data McMurdo

<table>
<thead>
<tr>
<th>Error Source</th>
<th>SZA = 60°</th>
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</thead>
<tbody>
<tr>
<td></td>
<td>310 nm</td>
<td>400 nm</td>
</tr>
<tr>
<td>Calibration, stability</td>
<td>2.7</td>
<td>2.1</td>
</tr>
<tr>
<td>Spectral resolution</td>
<td>0.8</td>
<td>0.0</td>
</tr>
<tr>
<td>(\lambda)-shift in UV-B</td>
<td>0.9</td>
<td>0.0</td>
</tr>
<tr>
<td>(\lambda)-shift + Fraunhofer lines</td>
<td>0.4</td>
<td>0.3</td>
</tr>
<tr>
<td>Nonlinearity</td>
<td>0.5</td>
<td>0.5</td>
</tr>
<tr>
<td>Stray light</td>
<td>0.0</td>
<td>0.0</td>
</tr>
<tr>
<td>Noise</td>
<td>0.7</td>
<td>0.4</td>
</tr>
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Extended uncertainty (\(k = 2\))

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<th>6.8</th>
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<td>3.4</td>
<td>7.9</td>
<td>3.1</td>
<td>3.4</td>
<td>3.4</td>
<td>3.4</td>
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</tbody>
</table>

Additional Uncertainties for Clear-Sky Conditions, Volumes 1–9

| Cosine error | 0.5 | 0.6 | 0.9 | 0.5 | 0.5 | 0.6 | 0.7 | 1.8 | 0.6 | 0.6 |
| Azimuthal error | 1.0 | 1.4 | 3.0 | 1.0 | 1.0 | 1.0 | 1.4 | 3.0 | 1.0 | 1.0 |
| Combined uncertainty | 3.3 | 2.7 | 3.8 | 3.0 | 3.2 | 3.3 | 2.7 | 4.1 | 3.0 | 3.3 |

Expanded uncertainty (\(k = 2\))

<table>
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<tr>
<th></th>
<th>6.6</th>
<th>5.4</th>
<th>7.6</th>
<th>6.0</th>
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<th>6.8</th>
<th>6.8</th>
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<td>3.9</td>
<td>3.0</td>
<td>3.3</td>
<td>3.3</td>
<td>2.9</td>
<td>5.5</td>
<td>3.0</td>
<td>3.3</td>
<td>3.3</td>
<td>3.3</td>
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</tbody>
</table>

Additional Uncertainties for Clear-Sky Conditions, Volumes 10–13

| Cosine error | 0.7 | 0.7 | 0.7 | 0.7 | 0.7 | 0.7 | 0.7 | 0.7 | 0.7 |
| Azimuthal error | 1.0 | 1.4 | 3.0 | 1.0 | 1.0 | 1.0 | 1.4 | 3.0 | 1.0 | 1.0 |
| Combined uncertainty | 3.2 | 2.8 | 3.9 | 3.0 | 3.3 | 3.3 | 2.9 | 5.5 | 3.0 | 3.3 |

Expanded uncertainty (\(k = 2\))

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<td>3.4</td>
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</tbody>
</table>

Additional Uncertainties for Overcast Conditions, Volumes 1–13

| Cosine error | 0.8 | 0.9 | 1.1 | 0.8 | 0.8 | 0.8 | 0.9 | 2.0 | 1.0 | 1.0 |
| Azimuthal error | 1.0 | 1.4 | 3.0 | 1.0 | 1.0 | 1.0 | 1.4 | 3.0 | 1.0 | 1.0 |
| Combined uncertainty | 3.4 | 2.8 | 3.9 | 3.1 | 3.3 | 3.4 | 2.8 | 4.2 | 3.1 | 3.4 |

Expanded uncertainty (\(k = 2\))

<table>
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<tr>
<td>Combined uncertainty</td>
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</tbody>
</table>

Additional Uncertainties for Overcast Conditions, Volumes 10–13

| Cosine error | 1.0 | 1.4 | 2.2 | 1.0 | 1.0 | 1.0 | 2.1 | 7.0 | 1.0 | 1.0 |
| Azimuthal error | 1.0 | 1.4 | 3.0 | 1.0 | 1.0 | 1.0 | 1.4 | 3.0 | 1.0 | 1.0 |
| Combined uncertainty | 3.4 | 3.0 | 4.3 | 3.1 | 3.4 | 3.4 | 3.4 | 7.9 | 3.1 | 3.4 |

Expanded uncertainty (\(k = 2\))

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<td>2.8</td>
<td>3.9</td>
<td>3.1</td>
<td>3.3</td>
<td>3.4</td>
<td>3.4</td>
<td>7.9</td>
<td>3.1</td>
<td>3.4</td>
</tr>
</tbody>
</table>

\(^a\)Ery, erythemal irradiance; DNA, DNA damaging irradiance. On the basis of individual uncertainties, a combined uncertainty is calculated and multiplied with a coverage factor of 2 in accordance with International Standards Organization [1993] guidelines. These final uncertainties are the quintessence of the uncertainty budget and are therefore printed in boldface.
Figure 2. Effective albedo calculated from SUV-100 spectra sampled during clear skies compared with reflectivity values from TOMS at McMurdo Station.

2.6. Calculation of Effective Albedo

Effective albedo is defined as the albedo of a uniform Lambertian surface, that, when used as input into a 1-D model, reproduces the measured spectrum [Lenoble et al., 2004]. Albedo leads to a wavelength-dependent increase in surface UV with larger changes at shorter wavelengths. The spectral effect is caused by the wavelength dependence of atmospheric backscatter [Lenoble, 1998]. In our implementation, we compare the average ratio \( r_{330} \) of measured and modeled UV irradiance in the wavelength interval 325–335 nm with the similar average ratio \( r_{400} \) for the interval 395–405 nm. The effective albedo value returned by the method is the model albedo value for which \( r_{330} = r_{400} \) becomes unity. The method is similar to “Method 2” presented in the works by Gröbner et al. [2000] and Weihs et al. [2001] and only applicable during clear-sky periods. Its accuracy depends mostly on (1) the relative calibration error of the two wavelength bands; (2) the uncertainty of the cosine error correction; and (3) the uncertainty of the model results, which in turn depends on the uncertainty of the ETS and other atmospheric processes that have a spectral effect, such as aerosol scattering. For example, changing aerosol optical depth from 0.025 to 0.035 in the model alters \( r \) by 0.3%. We estimate the 1σ-uncertainty of items 1 and 2 to 1.5% and 1%, respectively, on the basis of the uncertainty budget discussed in section 2.5. The uncertainty due to the model ETS is assumed to be 1.5%, and the combined 1σ-uncertainty of \( r \) is then 2.4%. A sensitivity analysis revealed that an error of 2.4% in \( r \) leads to an error in the retrieved albedo value of 0.11 when albedo is 0.6, and to 0.09 when albedo is 0.85. These uncertainties are similar to those reported by Weihs et al. [2001].

Figure 2 shows effective albedo values calculated from all available clear-sky spectra measured at MCM between December 1989 and January 2004 for SZA smaller than 85°. Albedo values range between 0.54 and 0.99 and tend to be highest in October and lowest in March. Statistics for every month are provided in Table 3. Figure 2 also shows reflectivity data extracted from Nimbus-7 and Earth Probe TOMS overpass data for times coinciding with SUV-100 albedo values. On average, TOMS reflectivity is larger by 0.04 ± 0.09 (±1σ) than SUV-100 albedo. Part of this bias may be due to the fact that TOMS reflectivity is the average over an area of 50 × 50 km², whereas SUV-100 albedo measurements are affected by the complex topography and surface condition in the vicinity of the instrument. There may also be occasions when the sky above the SUV-100 instrument is cloud free, but a significant fraction of the TOMS pixel is affected by clouds, resulting in an overestimate of surface reflectivity by TOMS.

[20] Albedo values used for modeling of version 2 spectra during cloudy periods were interpolated from the clear-sky data set. The interpolated values have larger (but difficult to quantify) uncertainties, in particular during February and March when albedo is most variable and clear-sky days are sparse.

2.7. Comparison of Version 0 and Version 2

[21] Version 2 data were compared with the originally released version 0 data. A detailed comparison of the two versions is available on the version 2 website (http://www.biospherical.com/nstf/Version2), focusing on seven spectral intervals in the UV and visible as well as erythemal [McKinlay and Diffey, 1987] and DNA-damaging [Setlow, 1974] irradiance. Differences are mostly caused by wavelength error and cosine error corrections.

[22] Below 310 nm, the wavelength error correction is an important factor because of the large increase of spectral irradiance in the ozone cut-off region of the solar spectrum. Differences between 300 and 310 nm range between −2% and +13%. In this wavelength range, the contribution of radiation from the solar beam contributes less than 25% to global irradiance at MCM. The cosine error correction is therefore dominated by the correction factor for diffuse skies, which is smaller than 1.073 for all years, and independent of SZA.

[23] Above 310 nm, the effect of wavelength errors is small and the difference of the two versions is almost entirely due to the cosine error correction. Differences between version 2 and 0 peak at SZAs between 72° and 84°, depending on the spectral band. Erythemal and DNA-damaging dose rates from the version 2 data set are between 0 and 9% higher than version 0 data. The differences depend on SZA, year, and sky condition, among other reasons. Differences for UV-A irradiance range between

<table>
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<th>N°</th>
<th>Avg°</th>
<th>Med°</th>
<th>Min°</th>
<th>Max°</th>
<th>Stdev°</th>
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<td>0.84</td>
<td>0.57</td>
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<tr>
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<td>0.81</td>
<td>0.68</td>
<td>0.99</td>
<td>0.07</td>
</tr>
<tr>
<td>Dec</td>
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<td>0.79</td>
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<td>0.07</td>
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<tr>
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<td>0.74</td>
<td>0.41</td>
<td>0.96</td>
<td>0.14</td>
</tr>
</tbody>
</table>

N°, number of data points; Avg, average; Med, median; Min, minimum; Max, maximum; Stdev, standard deviation.

*All, months September to March.
−5% and +12%. Differences for the integral between 400 and 600 nm range between −10% and +15% for SZA smaller than 70°. Differences peak between 78° and 84° and exceed 20% in some years. Ratios of version 2 and version 0 data show a different pattern before and after February 2000, when the instrument’s collector was modified.

2.8. Consistency of Version 2 Data

[24] After version 2 data were produced, the consistency of the new data set was examined by comparing clear-sky measurements of different years. In the first step of the analysis, the ratio \( q_{V2}(\lambda) \equiv \frac{E_{M}(\lambda)}{E_{C}(\lambda)} \) was calculated from all data associated with clear-sky periods. The comparison with the model is helpful to remove the dependence of known parameters such as SZA, total ozone, and albedo from the measurement, which are different every year. It also alleviates the difficulty that clear skies occur in different periods for every year. Note that model results were part of the correction procedures used to produce version 2 data. Total ozone and albedo were retrieved from measured spectra and used as model input parameters. Model results are therefore not independent from the measurement. Nonetheless, a comparison of measurement and model proved to be valuable in detecting problems of the measurements and in uncovering the influence of atmospheric processes on UV radiation, such as aerosol effects.

[25] In the second step of the analysis, the medians of \( q_{V2}(\lambda) \) were calculated for every month and year on a wavelength-by-wavelength basis, using all available ratio spectra \( q_{V2}(\lambda) \) with SZA smaller than 80°, in the selected month and year. The resulting median-ratio-spectra are denoted \( M(\lambda, m, y) \) and vary between 0.95 and 1.05 for wavelengths larger than 320 nm with larger deviations in some years. These ratios are typically not completely spectrally flat, but show a characteristic wavelength pattern, which is independent of year and month. A similar pattern was observed in \( M(\lambda, m, y) \)-ratios calculated for SPO [B04] and was associated with uncertainties of the ETS used in the radiative transfer model. Note that the patterns observed at MCM and SPO are slightly different because of the different ETSs used in this work and by B04.

[26] For the final step of the analysis, we consider the ratio \( Q(\lambda, p, y) \), defined as:

\[
Q(\lambda, p, y) = \frac{M(\lambda, m, y)}{\frac{1}{11} \sum_{y=1994}^{2004} M(\lambda, m, y')}
\]

The average in the denominator of equation (1) includes only years with background (or near-background) aerosol conditions. Because of the construction of \( Q(\lambda, m, y) \), systematic differences between measurement and model, which affect all years equally such as those arising from the ETS, are ratioed out. These “\( Q \)-ratios” are shown in Figure 3, which consists of 6 plots for the months October to March.

[27] With the exceptions of the cases explained below, \( Q \)-ratios of all years and periods are within the range 1 ± 0.05. Corrected measurements of different years are therefore consistent to within ±5% and agree within the measurement uncertainty (Table 2).

[28] \( Q \)-ratios for November 1991 and January to March 1992 tend to be low, particularly at wavelengths in the visible. This is likely due to the presence of volcanic
aerosols during this period and will be discussed in section 3. Note that aerosols may have also affected radiation levels in October 1991. Data of this month could not be included in Figure 3 because of the lack of cloud-free periods.

[39] \( Q \)-ratios for the period December 1989 and February 1990 are high by 5–10\%. This period marks the start of operational measurements at MCM. At that time, neither calibration procedures nor calibration standards were as well established as when compared to later years of operation. The cosine error correction is also uncertain during this period.

[40] \( Q \)-ratios of several years show distinct humps or dips between 480 and 525 nm, which is related to a Wood anomaly [Palmer, 2002; B04] of the SUV-100’s monochromator. In some years, the cosine error correction algorithm was less successful in removing this artifact than in others. \( Q \)-ratios for November 1994 are relatively low in the UV. During this period, the model albedo was calculated to a relatively high value of 0.90, which may have been too high.

### 3. Effect of Volcanic Aerosols on UV

[41] Figure 3 indicated that clear-sky measurements of November 1991 and January to March 1992 were abnormally low compared to measurements of other years. Here we provide evidence that this could be caused by attenuation of solar radiation by aerosols from the Mount Pinatubo and Cerro Hudson eruptions, which occurred in June and August of 1991, respectively. The Antarctic lower stratosphere in the spring of 1991 was characterized by a layer of volcanic aerosol from the Cerro Hudson eruption. After 20 September 1991, the aerosol was observed daily at McMurdo at altitudes between 9 and 13 km by Lidar [Deshler et al., 1992]. Between October 1991 and January 1992, monthly average aerosol optical depth anomalies in the 530–690 nm range varied between 0.21 and 0.275 at SPO, according to measurements performed by CMDL with wideband filter wheel normal incidence pyrheliometers [Dutton and Christy, 1992; CMDL, 2002]. In order to investigate whether the low ratios in Figure 3 can be explained by the effect of aerosols, we compared our measurements with version 2 model spectra (which are based on background aerosol conditions), and with a second set of model spectra, which took volcanic aerosols into account. The layer of volcanic aerosols was assumed to be between 9 and 13 km. The aerosol optical depth at 550 nm was set to 0.25, and single scattering albedo was set to 0.98. The wavelength dependence of aerosol extinction was parameterized with Angström’s turbidity formula by setting the Angström coefficient \( \alpha \) to 0.9. The value of \( \alpha \) was estimated with Mie calculations from the aerosol size distribution measured at McMurdo by Deshler et al. [1992] on 8 October 1991.

[42] Spectra measured under cloud-free conditions between October and December 1991 were selected from the version 2 data set and compared with version 2 model spectra. The resulting ratio-spectra were binned into 5° SZA-bands. Medians of these ratio-spectra were calculated for each bin and are presented in Figure 4a. Figure 4a indicates that the deviation of measurement and model increases with increasing SZA and increasing wavelength. This pattern is quantitatively consistent with the effect of aerosols described by B04 for SPO data. The analysis was repeated by using the second set of model spectra that take volcanic aerosols into account. The resulting median-spectra are virtually independent of SZA and wavelength (Figure 4b). The bias between measurement and model of approximately 4\% seen in Figure 4b is within the range of discrepancies seen for other years (Figure 3), and within the uncertainties of measurement and model. The dip at 575 nm is caused by absorption of the \( \text{O}_2 \)-\( \text{O} \) collision complex [Greenblatt et al., 1990], which is not considered in the model. Figure 4 indicates that the low clear-sky measurements in 1991 can indeed be explained by aerosol attenuation.

### 4. UV Climatology at McMurdo Station

#### 4.1. Time Series and Trends

[43] After version 2 data for MCM had been generated, the data set was used to establish a climatology of ultraviolet and visible radiation. Figure 5 shows time series of daily doses of four version 2 data products for the period December 1989 to January 2004. Daily doses were calculated by integrating instantaneous measurements of the four data products over 24 hour periods. Gaps in time series of 2 hours or less were filled via spline interpolation. Days with gaps longer than 2 hours were excluded from further analysis. The four data products are spectral irradiance at 305 nm, erythemal irradiance (action spectrum by McKinlay and Diffey [1987]), spectral irradiance integrated between 342.5 and 347.5 nm, and spectral irradiance integrated between 400 and 600 nm. Figure 5 also shows daily doses of short-wave irradiance (0.3–3.0 \( \mu \text{m} \)) measured by NIWA.
with a pyranometer at New Zealand’s research station Scott Base, which is located at sea level 4 km southeast of Arrival Heights. Daily doses measured in different years on the same calendar day were averaged to setup a climatology. These climatological mean values are indicated by thin lines in Figure 5 for all five data products.

Figure 5 supports the following conclusions:

1. In some years, daily doses of the spectral irradiance at 305 nm show large deviations from the climatological mean, in particular during the months September to December, which are affected by the ozone hole. Years with substantially enhanced doses were 1990, 1992, 1998 (overall maximum), 1999, and 2001. Years with comparatively low doses were 1991, 1994, 2000, and 2002.

2. Daily erythemal doses show a similar pattern as the 305-nm dose, but variability and relative departures from the climatological mean are generally smaller. This can be expected since erythemal irradiance is less sensitive to changes in atmospheric ozone amounts than spectral irradiance at 305 nm. The largest erythemal daily dose was 6.7 kJ/m², measured on 28 November 1998.

3. Daily doses calculated from the integral 342.5–347.5 nm show a similar pattern in every year with little departure from the climatological mean. This can be expected as this integral is not affected by total ozone. Closer inspection reveals subtle but significant difference between data of the different years. For example, data from February and March tend to be below the mean between 1990 and 1993, and above the mean between 2001 and 2003.

4. Daily doses calculated from the integral 400–600 nm exhibit a similar behavior to daily doses of the 342.5–347.5 nm integral. February and March values are also below the mean for the years 1990 and 1993, and tend to be slightly above the mean in later years. Short-term variability is slightly larger than for the 342.5–347.5 nm integral because of the larger influence of clouds at larger wavelengths [Seckmeyer et al., 1996; B04].

5. Daily doses calculated from pyranometer measurements at Scott Base exhibit a similar pattern than the 400–600 nm integral, but the day-to-day variability is generally larger. Also this data set indicates that doses from February and March 1990–1993 are below the mean. Since this data set is completely independent from SUV-100 data, the observed differences are likely real and cannot be explained with possible (though unlikely) instrumental drifts in the SUV-100 data set, or very localized changes, for example caused by changes in albedo in the immediate vicinity of the SUV-100.

In order to quantify long-term changes, we calculated monthly averages of the daily doses depicted in Figure 5. Only months with at least 25 daily dose values were used for further analysis. UV data measured between December 1989 and March 1990 were not included because of their...
larger uncertainty. Figure 6 shows the resulting time series of monthly average daily doses of the four SUV-100 data products and the short-wave data set. Trend lines were calculated for all data sets by linear regression and are also shown in Figure 6. Slopes of the trend lines (expressed in change per decade relative to the year 1990), their 2σ-uncertainty, and regression coefficients \( R^2 \) are given in the legend of Figure 6.

[41] Average monthly doses calculated from spectral irradiance at 305 nm and erythral irradiance exhibit large year-to-year variability during September and December as atmospheric ozone concentrations fluctuate substantially from year to year. Trends vary between −36% and +12% per decade and are positive in September, November and December, and negative in October. However, none of these trends are significant at the 2σ level. Clearly, the large interannual variability in ozone obscures any linear trend, if it exists. Monthly average doses for November are larger than December doses in 1998 and 2001, indicating that the effect of the smaller SZA in December is offset by the lower total ozone values in November.

[42] Monthly average doses for the 342.5–347.5 nm and 400–600 nm integrals show little variability from year to year. Trends for the months September to December are less than ±1% per decade and not statistically significant. This indicates that cloudiness and albedo have likely not changed during the time period considered. Monthly average doses for October 1991 are somewhat low, possibly because of influence of volcanic aerosols (section 3). Dose values during the austral spring of 1997 were unusually low, too. The reason is possibly unusually low surface albedo during this year. Monthly average short-wave doses measured at Scott Base show a similar interannual pattern to that of the 400–600 nm integral. For example, data from 1997 are low in both data sets; data from November 2001 are high.

[43] Daily doses for January are relatively sparse because the annual instrument service normally takes place during this month. Therefore no conclusions can be drawn. For the months February and March, monthly average doses of all five data products exhibit statistically significant increases, which range between 11% (short-wave dose in February) and 34% (305 nm dose in February) per decade. Trends are generally decreasing with increasing wavelength and are smallest for the short-wave dose. We confirmed with further analysis that these increases cannot be an artifact of the distribution of missing days when calculating the monthly average. Significant positive trends (although of somewhat smaller magnitude) were also apparent in version 0 data, ruling out the possibility that erroneous corrections applied during processing of version 2 may have lead to a time-dependent bias. We therefore conclude that the observed increases are real. Further analysis presented below suggests that the most likely reasons for these trends are changes in cloudiness and/or surface albedo.

[44] A similar analysis based on noon-time irradiance instead of daily dose is available at the version 2 website (http://www.biospherical.com/nsf/Version2). Observed trends for noon-time irradiance are very similar to those found in daily doses. Typical noon-time UV Indices in November and December vary between 2 and 5.5, but larger values have been observed when low total ozone amounts occur late in the year when prevailing SZAs are small. The largest UV Index was 7.5 and was measured on 28 November 1998. (UV Index is an internationally recognized measure of sun-burning UV irradiance, and is defined as spectral irradiance weighted with the CIE erythral action spectrum [McKinlay and Diffey, 1987], expressed in the units of W/m², and multiplied by 40 [World Meteorological Organisation (WMO), 1998].)

4.2. Effect of Clouds and Albedo on UV

[45] Clouds are the dominant factor in reducing UV radiation below clear-sky level. In a high-albedo environment, the effect of clouds is substantially modified by surface albedo [Nichol et al., 2003; Ricchiazzi et al., 1995]. The effects of clouds and the modification of cloud attenuation by surface albedo are therefore discussed together.

[46] The modification of UV radiation by the combined effects of clouds and albedo was quantified with transmittance \( T(t) \), defined as

\[ T(t) = \frac{E_{M,345}(t)}{E_{C,345}(t)} \times \sum_{i \in \{P(t),CS\}} \frac{1}{n} E_{M,345}(t) \]  

\( E_{M,345}(t) \) is the measured version 2 342.5–347.5 nm integral at time \( t \), and \( E_{C,345}(t) \) is the associated clear-sky model value. The denominator of equation (2) is the average ratio of measurement and model for clear-sky conditions (CS) during year \( y \) and a two-week period \( P(t) \) (1–15 January, 16–31 January, . . . , 16–31 December). This term corrects for the observed bias between measurement and model for clear skies in any given year and period. Corrections vary between 0.93 and 1.07; the average correction is 0.98. Albedo used for the calculation of \( E_{C,345}(t) \) was determined from clear-sky spectra as described in section 2.6.

[47] In the next step, \( T(t) \)-values for the period 1–15 January were selected from data of the years 1989–1996, and binned into 0.02-wide intervals to set up a frequency distribution. Similar distributions were constructed for each period \( P(t) \) and are shown in Figure 7. Each plot in Figure 7 represents the climatology of transmittance relative to clear-sky for a different part of the year. A similar analysis was also performed for data of the years 1997–2004 and is shown in Figure 8.

[48] Distributions of most periods display a distinct maximum at \( T(t) = 1 \), marking clear-sky conditions. Some distributions have a secondary maximum near 0.9, which is likely related to situations when the direct beam of the Sun is blocked by clouds with fair weather conditions otherwise. Minimum ratios may be as low as 0.3. Enhancement of radiation beyond the clear-sky value rarely exceed 10%. Frequency distributions for the years 1997–2004 are generally smoother than those for the years 1989–1996 because of the larger number of spectra measured in later years. However, there are significant differences between the two periods. Specifically, distributions for February and March of the period 1989–1996 have a smaller clear-sky fraction and are generally shifted to lower values compared to distributions of the same months for the 1997–2004 time frame. This is also indicated in the average \( T(t) \)-values, noted in the legends of these plots, which are lower by 3–12% in February and March of the earlier period. This
Figure 6. Time series of average monthly daily dose for the months of (left) January to March and (right) September to December for five data products, indicated in the top left corner of each plot. Dashed lines are trend lines determined by linear regression. Trend per decade relative to 1990, 2σ-uncertainty of trend, and regression coefficient $R^2$ are also indicated.
demonstrates again that cloud and/or albedo conditions have changed between the two periods. Frequency distributions for other months are similar for the two time frames. It should be noted that the years 1991–1994 were affected by volcanic aerosols, whereas the aerosols were near background levels between 1995–2004. However, this can only explain a part of the observed differences since aerosols also affected radiation levels during the months September to January when distributions of the two periods are similar. Maximum reductions of $T(t)$-values that can be expected from aerosol extinction are also much smaller than those shown in Figures 7 and 8. A similar analysis for the 400–600 nm integral can be found on the version 2 website (http://www.biospherical.com/nsf/Version2). These distributions are qualitatively similar to the distributions for the 342.5–347.5 nm integral, but confirm that the effect of clouds is larger in the visible than for the UV: average $T(t)$-values may be as low as 0.678. Some rationale for the increased effect of clouds at longer wavelengths was presented by B04.

[49] February and March are the months with the greatest variability in albedo (Figure 2). A long-term change in surface albedo (e.g., caused by possible changes in sea ice extent and snow fall) could therefore be one of the reasons for the observed change. Unfortunately, clear-sky days are very rare in February and March and surface

Figure 7. Frequency distributions of transmittance $T(t)$ defined as the ratio of spectral irradiance integrated over 342.5–347.5 nm to the associated clear-sky irradiance, calculated from measurements of the years 1989–1996. Each of the 15 plots refers to a different two-week period as indicated in the top left corner of each plot. SZA range, number of data points $N$, width of the histogram columns (Bin), average (Avg), and standard deviation ($\sigma$) of the distributions are also provided.

Figure 8. Same as Figure 7 but for period 1997–2004.
albedo can be calculated only on few occasions. It is not possible to determine trends in albedo for these months with confidence.

[50] Ricchiazzi et al. [1995] have suggested a method to decouple the effects of clouds and albedo using measurements of global spectral irradiance at 410 and 630 nm. The method could also be applied to our data. However, Ricchiazzi et al. [1995] point out that their method is very sensitive to calibration errors. For example, they estimate that differences in cloud optical depth (COD) of a factor of 3–4 may result from a calibration uncertainty of only ±2%. They conclude that the retrieval technique is probably not appropriate to make precise measurements of COD, or to assess small variations of albedo such as might result form aging snow.

[51] We also note that effective albedo derived from clear-sky spectra may not be applicable to cloudy situations because of multiple reflections between the surface and the cloud base. Multiple reflections change the effective surface area (or radius around the instrument) that affects UV levels at the place of the instrument [Degünther and Meerkötter, 2000]. The fact that clouds are usually not homogeneous (as it is required in the method by Ricchiazzi et al. [1995]) leads to further complications.

[52] Because of these limitations, we did not attempt to simultaneously retrieve COD and effective surface albedo from our measurements during periods of cloud cover. Instead, we performed model calculations of \( T(t) \) as a function of COD and effective albedo allowing to estimate the expected magnitude of changes in UV related to changes in one or both of the two parameters. For these calculations we assumed that clouds consist of a homogeneous and stratiform layer located between 2 and 3 km altitude. Location of the cloud layer was estimated from cloud ceiling data observed at McMurdo in 1997 and provided by the National Climatic Data Center (NCDC). The median cloud ceiling height was 1500 m. The effective water droplet radius was set to 10.0 μm. Calculations mentioned by Ricchiazzi et al. [1995] suggest that calculated CODs are rather insensitive to cloud height and droplet size. Atmospheric pressure was set to 966 hPa and aerosol attenuation was parameterized as described in section 2.2

[53] Figure 9 presents transmittance at 345 nm and SZA = 60° as a function of COD for various values of surface albedo. Figure 9 indicates for example that a change in transmittance from 0.8 to 0.75 could be caused by a change in albedo from 0.80 to 0.69 at a constant COD of 9.1 or a change of COD from 9.1 to 13.3 at a constant albedo of 0.80. Similar calculations at SZAs of 50°, 70° and 80° lead to only slightly different results. Calculations at λ = 500 nm confirmed that changes in transmittance are larger at longer wavelengths, both for changes in cloudiness and/or albedo. It is important to note that the spectral effect is too small to reliably quantify the fractional contributions of either of the two parameters to the differences seen between Figures 7 and 8.

4.3. Variability in UV Due to Variations in Total Ozone

[54] The influence of total ozone on UV depends strongly on wavelength and therefore has to be evaluated separately for each wavelength or UV data product of interest. Here we focus on the UV Index, denoted \( E_{305}^M(t) \), and spectral irradiance at 305 nm, \( E_{305}^M(t) \).

[55] To reduce the effect of clouds, we divided both quantities by \( E_{345}^M(t) \). Note that the division with \( E_{345}^M(t) \) does not remove the effect of clouds completely since attenuation by clouds is not entirely independent of wavelength. This effect is discussed in more detail by B04.

\[
\tilde{E}_{M,UVI}(t) = E_{M,UVI}(t)/E_{M,345}(t)
\]

To quantify the year-to-year variability of \( \tilde{E}_{M,UVI}(t) \), we selected data from different years that were measured at the same day and hour, and calculated their average. The ratio of individual measurements to this average is then used to describe interannual variability. The ratio is denoted \( V_{UVI}(t) \), and formally defined as:

\[
V_{UVI}(t, y) = \frac{1}{\pi} \sum_{y' \in [1999, 2004]} \tilde{E}_{M,UVI}(t, y')
\]

where \( t \) is time within a given year \( y \). \( V_{UVI}(t) \) values are used for further analysis only, if at least data from 8 years contribute to the average.

[56] Similar to the approach described in section 4.2., we constructed frequency distributions for \( V_{UVI}(t) \) using data from all years and 0.05-wide bins. The distributions are presented in Figure 10. Distributions for January to April are much narrower than distributions for October to December. During the first half of the year, less than 1.5% (2.3% for period 16–31 March) of \( V_{UVI}(t) \) values differ by more than ±17.5% from unity. Standard deviations of the distributions are less than 0.079. For the period 1 October to 15 December, \( V_{UVI}(t) \) values differ by up to 93% from unity, and the standard deviations for October and November are approximately three times as large as during the first half of the year. The broader distributions are mainly caused by the influence of the ozone hole, which typically disappears sometime during the period of late November to the beginning of December [WMO, 2003].
A comparison of Figures 7, 8 and 10 (and in particular the standard deviations of the distributions) suggests that during the first half of the year, the variability in UV introduced by clouds and albedo is about twice as high as that introduced by ozone. Between September and November, the situation is reversed and the variability due to ozone is about double that due to clouds.

Frequency distributions for $V_{305}(t)$ are shown in Figure 11. Spectral irradiance at 305 nm is more strongly affected by ozone variations than the UV Index. Distributions for $V_{305}(t)$ are therefore much broader than for $V_{UVI}(t)$; standard deviations for the same months are between a factor of 2.2 and 3.8 larger for the $V_{305}(t)$ distributions. A similar difference was reported by B04 for SPO. During the first half of the year, $V_{305}(t)$ varies between 0.5 and 1.5. During the period affected by the ozone hole, $V_{305}(t)$ can be as low as 0.08 and as high as 4.3, suggesting that in isolated events, spectral irradiance at 305 nm can be higher by a factor of 50 in some years than in others. Distributions between 16 October and 31 November have their maximum at low $V_{305}(t)$-values, which occur when MCM is little affected by the ozone hole. The long tail toward larger values is due to the large variability during periods affected by the ozone hole. During January to March, the variability in UV introduced by clouds and albedo is comparable to that introduced by ozone.

Between September and November, the variability due to ozone is 4–6.5 times as high as that due to clouds.

5. Discussion and Conclusions

Version 2 data from the NSF/OPP UV spectroradiometer at MCM have been produced and feature smaller uncertainties and a larger number of data products than the original release “version 0.” New products include (but are not limited to) total ozone column and effective albedo for clear skies. Corrections applied to version 2 data were adopted from the work of B04, which also discusses the assumptions and limitations of the procedure in detail. A comprehensive uncertainty budget was established, and the new data set was compared to version 0 data and results of a radiative transfer model to further assess its quality and homogeneity. Erythemal and DNA-damaging dose rates from the version 2 data set are higher by 0–9% than for version 0 data, but differences for wavelengths in the visible can be larger. Clear-sky measurements of different years typically agree to within ±5%, with some exceptions discussed in section 2.8. Most notably, clear-sky measurements between November 1991 and March 1992 are smaller than in other years, and we showed that these low readings can be explained by the presence of volcanic aerosols. The magnitude of the aerosol effect is consistent with that observed at SPO [B04]. Lastly, a climatology of UV radiation at MCM was established, focusing on changes in UV observed over time, the combined effect of clouds and aerosols, and the variability introduced by total ozone fluctuations.

Daily doses of spectral irradiance at 305 nm and daily erythemal doses are largest in November when low ozone columns coincide with relatively small SZAs. Linear regression analyses of monthly average daily doses did not indicate statistically significant trends for the months of September through January. Highest values were observed in the austral spring of 1998 and 2001 and comparatively low values in 2000 and 2002. This is consistent with measurements at SPO [B04] and also with emerging evidence that ozone depletion may have reached its peak in the 1990 and first years of the 21st century, and is now starting
to recover [Newchurch et al., 2003]. However, it is too early to draw final conclusions considering that only the last 2–3 years showed comparatively low UV levels.

Doses in the UV and visible from February and March are significantly lower during the first part of the time series (1989–1996) than during the second (1997–2004). This is likely caused by changes in cloudiness and albedo, or both. Herman et al. [2001] have observed large increases in reflectivity between 1980 and 1992 around Antarctica from the Nimbus-7 TOMS satellite data, and attributed those changes to increases of cloud reflectivity. By using additional data from the Earth Probe TOMS satellite, the record was extended to the year 2000, indicating that the reflectivity increases have continued, but at a slower rate [Herman, 2003]. This conflicts with our observations, which suggest that clouds have decreased. However, it is difficult to assess changes in cloudiness from TOMS data at the location of our UV spectroradiometer because of its location at the Antarctic coast and complex topography.

In March 2000, a very large (285 × 37 km²) iceberg (labeled B-15 by the National Ice Center), calved from the eastern portion of the Ross Ice Shelf [Arrigo et al., 2002]. As of 2005, fractures of the iceberg still affect weather systems and ocean currents near Ross Island. This led to severe ice conditions during recent years at McMurdo Sound, and possibly to an increase in albedo, which could explain some of the trend seen in UV data.

Determining the relative portion of changes in cloud cover and albedo to changes in UV with sufficient accuracy would require measurement uncertainties of less than ~2% for wavelength in the UV and visible. Unfortunately, the uncertainty of our measurements is higher (Table 2), preventing definitive conclusions. We also note that our trend estimates are larger in the UV than the visible, which cannot be explained with decreases of COD or increases of effective albedo alone. An explanation of the observed wavelength dependence may require 3-dimensional modeling, taking into account the complex topography at Arrival Heights. For example, Deguinther and Meekotter [2000] showed that an area of high surface albedo located several kilometers away from the measurement site has a smaller influence at 300 nm than at 330 nm. A more detailed analysis is beyond the scope of this paper.

On average, clouds reduce UV radiation at 345 nm by 10% of the clear-sky value. Reductions vary by month and year, can exceed 60% on rare occasions (Figures 7 and 8), and generally increase with wavelength. Cloud effects at MCM are larger than those observed at SPO [B04]. The comparatively little influence of clouds at SPO is due to higher surface albedo and the lower atmospheric water content, which leads to optically thinner clouds than observed at the coast [Lubin et al., 1992].

Time series of daily doses of spectral irradiance at 305 nm and of erythemal daily dose demonstrated the large interannual variability during the austral spring. The variability is caused by annual difference in the depletion depth and timing of the ozone hole and has been quantified by calculating departures from the climatological mean (Figures 10 and 11). Between September and November, the variability due to ozone is 4–6.5 times as high as that due to clouds.

Acknowledgments. SUV-100 measurements at McMurdo were supported by the National Science Foundation’s Office of Polar Programs via subcontracts to Biospherical Instruments Inc. from Antarctic Support Associates (ASA) and Raytheon Polar Services Company (RPSC). G. Harris from Research Instrument Systems designed and built the first prototype of the SUV-100. We wish to express our gratitude to numerous operators of the SUV-100 spectroradiometer at MCM. We further thank T. Deshler from the University of Wyoming for the ozone sonde measurements from McMurdo and J. Wild from the Network for the Detection of Stratospheric Change (NDSC) for making the most recent data available prior to their public release. All data older than two years are available on the NDSC database (http://www.ndsc.ws). Atmospheric pressure data from McMurdo were provided by Kathie Hill, RPSC, and Matthew Lazzara, University of Wisconsin-Madison. Further thanks go to R.D. McPeters from NASA for providing TOMS total ozone data.

References


G. Bernhard, C. R. Booth, and J. C. Ehramjian, Biospherical Instruments Inc., 5340 Riley Street, San Diego, CA 92110, USA. (bernhard@biospherical.com)