7.8. Differences Between Sites

There are substantial differences in the UV climatology between the various sites of this network. A significant portion of the differences can be traced to their geographical locations, as lower latitudes experience higher sun elevations and more UV, all other factors being equal. However, the contention that insignificant levels of UV will occur in polar regions because of latitudinal differences is not true. In the following discussion, measurements from all six network sites are presented. The comparison is based on the average and maximum UV Index[+], as well as average and maximum daily dose. The results reveal that the observed differences between the sites very much depend upon the selection of the quantity used for the comparison.

Average daily dose is an appropriate quantity for investigating the cumulative UV exposures received over an extended time-period. It does not capture, however, the impact of transient high levels of UV that may occur during episodic combinations of clear skies and severe ozone depletion. Such incidents may have biological significance in systems that do not obey reciprocity in terms of exposure intensity versus duration. To study those effects, maximum daily dose is the more suitable measure. For some organisms high levels of UV radiation received during short time-periods, ranging from minutes to hours, may be even more detrimental. This also includes human beings who are usually exposed for few hours rather an entire 24-hour period. Since the highest levels of UV radiation are normally occurring around solar noon, the average and maximum noontime UV Index are suitable quantities for studying short-term UV effects on humans.

Figure 7.8.1 shows average daily erythemal dose for all sites but Summit. (Data from Summit were excluded because they cover less than one year.) Summer doses in San Diego are highest because of the higher solar elevation at this low-latitude site. However, average November UV doses at McMurdo, Palmer Station, and South Pole can reach 80-100% of typical mid-summer San Diego conditions. Note that average daily erythemal doses at McMurdo and South Pole Station are very similar between January and April but disagree significantly in November when doses at the South Pole exceed McMurdo levels up to 30%. The reason is that the influence of the ozone hole on UV-levels is more pronounced at the South Pole than at the Antarctic coast, and the solar elevation is constant for 24 hours. Radiation levels at Ushuaia between November and January reach about 70% of San Diego values. Average daily erythemal dose at Barrow is the lowest of all network sites because of the high-latitude location of Barrow and the fact that ozone depletion in the northern hemisphere is less severe than over Antarctica.

All dose values discussed above represent irradiance values integrated over a period of 24 hours. Since McMurdo and South Pole have 24 hours of sunlight during summer, there is a significant contribution to the daily dose from the “midnight sun,” which diminishes the difference to a site outside the Polar Circle. The additional contribution from the midnight sun is missing when noontime UV Indices are compared, leading to larger inter-site differences. Figure 7.8.2 proves that differences between sites are indeed larger when average noontime UV Indices rather than daily erythemal doses are compared. The average noontime UV Index during summer at San Diego typically ranges between 8 and 10. During the austral summer, average noontime UV Indices at Palmer Station and Ushuaia are about 55% of the San Diego value. This difference is clearly higher compared to the difference seen for daily doses. The difference between sites further increases with their latitude differences: Average noontime UV Indices at McMurdo and the South Pole are about 4 and 2, respectively.

The differences between sites show a completely different pattern when maximum rather than average values are compared. Figure 7.8.3 shows the maximum UV Index observed during a given day of the year between 1991 and 2005. At Palmer Station, the maximum observed UV Index was 14. This value is about 25% higher than the highest UV Index measurement at San Diego. The particularly high maximum levels at Palmer Station can be observed when the ozone hole starts to dissolve in November and December.

[+] UV Index is an internationally recognized measure of sun-burning UV irradiance, and is defined as spectral irradiance weighted with the CIE erythemal action spectrum, expressed in the units of W/m², and multiplied by 40. UV Index is identical with the values of the published Database 3 data product “Dose3_CIE_Erythema” multiplied with 0.4. See Section 6 for further explanation.

During this part of the year, the polar vortex becomes unstable and air masses with low ozone concentration may be centered over Palmer Station. In combination with the relatively high solar elevation, this leads to noontime UV levels that exceed San Diego summer levels. The maximum UV Index at Ushuaia was 11, which is similar to summer-time values at San Diego. Although maximum UV Indices at McMurdo, South Pole and Barrow are significantly smaller than UV Index values at San Diego, the difference is clearly much smaller in comparison to average noontime values.

The picture changes again when we compare maximum daily doses rather than instantaneous UV Indices. Figure 7.8.4 shows that maximum daily erythemal doses at Palmer Station are up to 33% higher than in San Diego. Maximum doses at McMurdo, South Pole Station, and Ushuaia are comparable to levels in San Diego. Note that the difference between McMurdo and the South Pole is also much smaller when maximum daily doses rather than maximum noontime UV Indices are compared. We attribute this to the difference in the diurnal cycle of the sun at these sites. At McMurdo, radiation levels peak at local solar noon whereas at the South Pole, there is virtually no change in solar elevation during one day.

The difference in maximum and average daily erythemal doses can be seen by comparing Figure 7.8.1 and Figure 7.8.4. The comparison reveals a 20% difference for San Diego and a 25% difference for Barrow. Maximum daily doses at the four austral high-latitude sites are roughly 50-100% higher than average daily doses, reflecting again the important influence of ozone variability on UV at southern high latitudes.

All observations above were based on erythemally weighted quantities. In contrast, Figure 7.8.5 shows a comparison between the sites based on average daily DNA-dose. Although the pattern presented in this figure is similar to Figure 7.8.1, there are also important differences. Since DNA-weighted irradiance is about a factor of two more sensitive to changes in ozone than erythemal irradiance (see Section 7.10 and in particular Figure 7.10.3.), ozone related features are more pronounced in Figure 7.8.5. For example, the peak in the DNA-weighted data for Palmer Station in November is much less apparent in erythemal data. Similarly, average DNA-dose observed at the South Pole Station drops by about 40% during end of November whereas the drop in erythemal doses is only 20%. A comparison of Figure 7.8.1 and Figure 7.8.5 further reveals that summer-time erythemal doses from the high-latitude sites and from San Diego are more comparable than DNA-weighted doses. The reason is that erythemally weighted irradiance depends less on solar zenith angle than DNA irradiance. Therefore, the difference in latitude between the sites (which is directly linked to the difference in solar zenith angle) has a smaller impact on erythemal doses.

Figure 7.8.6 presents average daily irradiation for the 400-600 nm band from all sites. Since spectral irradiance in this band depends only very little on atmospheric ozone concentrations, there are no ozone-related features in this figure; the curves appear smoother than for DNA and erythemal dose. Note that summer doses of the 400-600 nm interval are highest at the South Pole and McMurdo (the network sites with the highest latitudes), exceeding doses in San Diego. This is caused by 24 hours of sunlight, high surface albedo, and for South Pole, high altitude (see also Section 7.10).

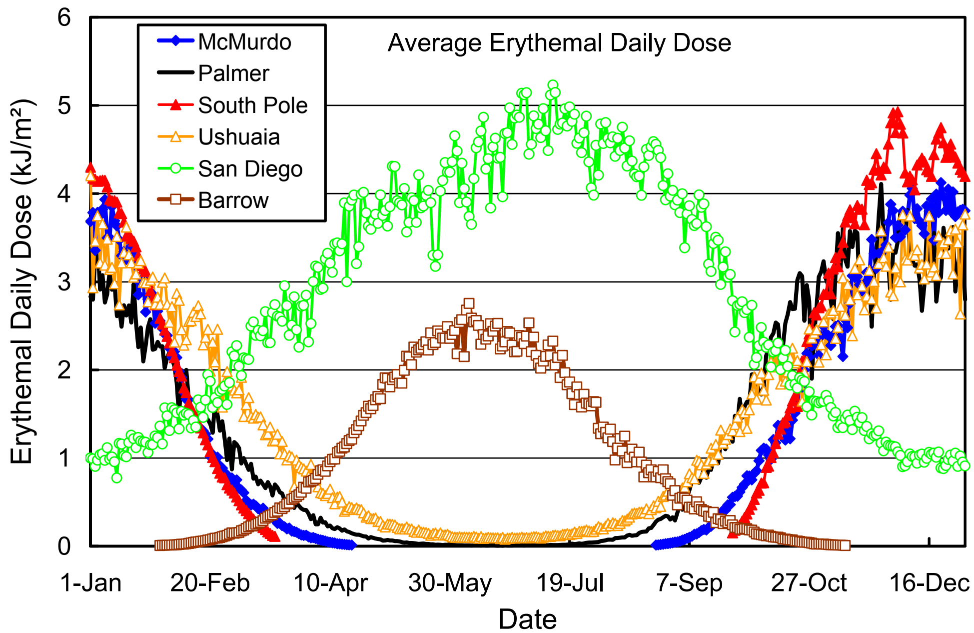

Figure 7.8.1. *Comparison of average daily erythemal dose from all network sites. The data is based on average daily doses from the years 1991 – 2004.*

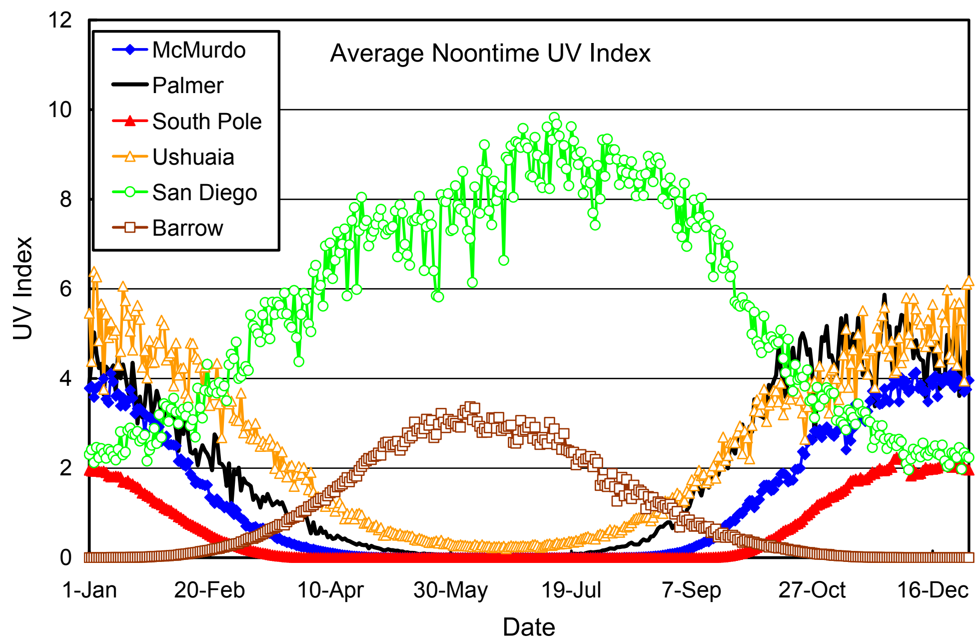

Figure 7.8.2. *Comparison of average noontime UV Index from all network sites. Daily measurements from the years 1991 – 2004 have been averaged.*

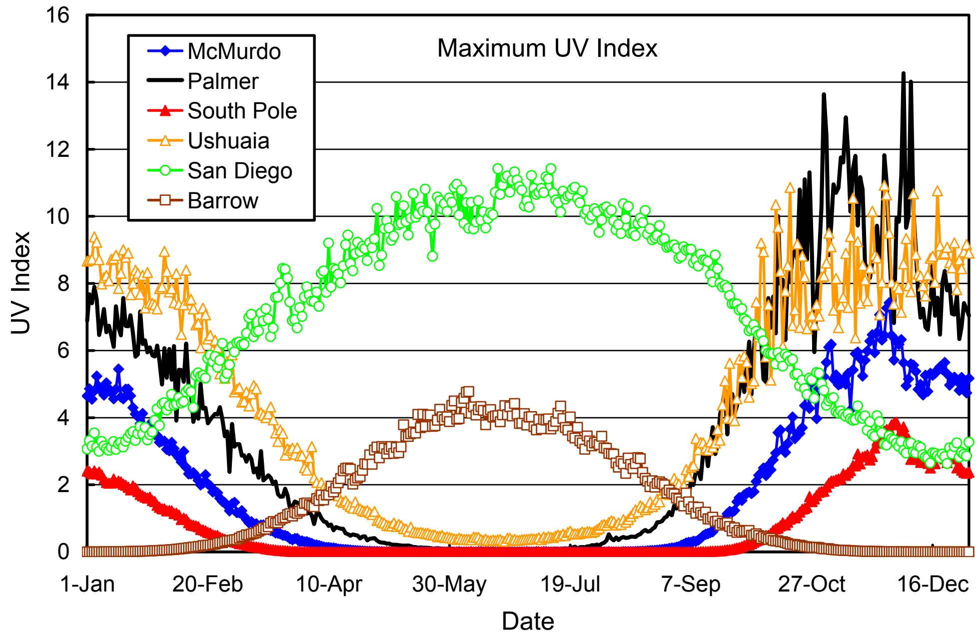

Figure 7.8.3. *Comparison of maximum UV Index from all network sites observed at a given day of the year. The data is based on measurements from the years 1991 – 2004.*

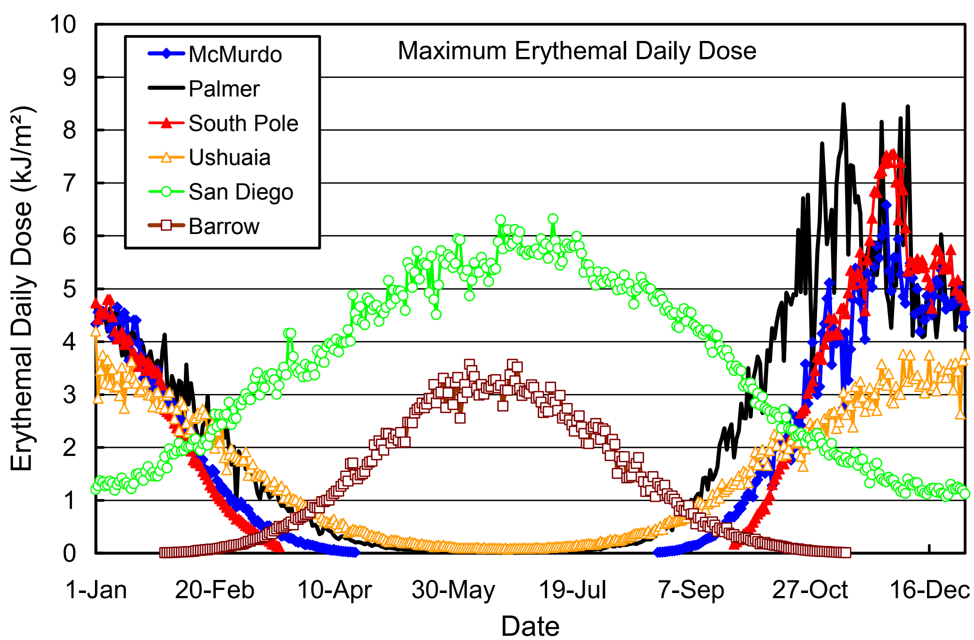

Figure 7.8.4. *Comparison of maximum erythemal daily dose from all network sites. The data is based on daily doses of the years 1991 – 2004.*

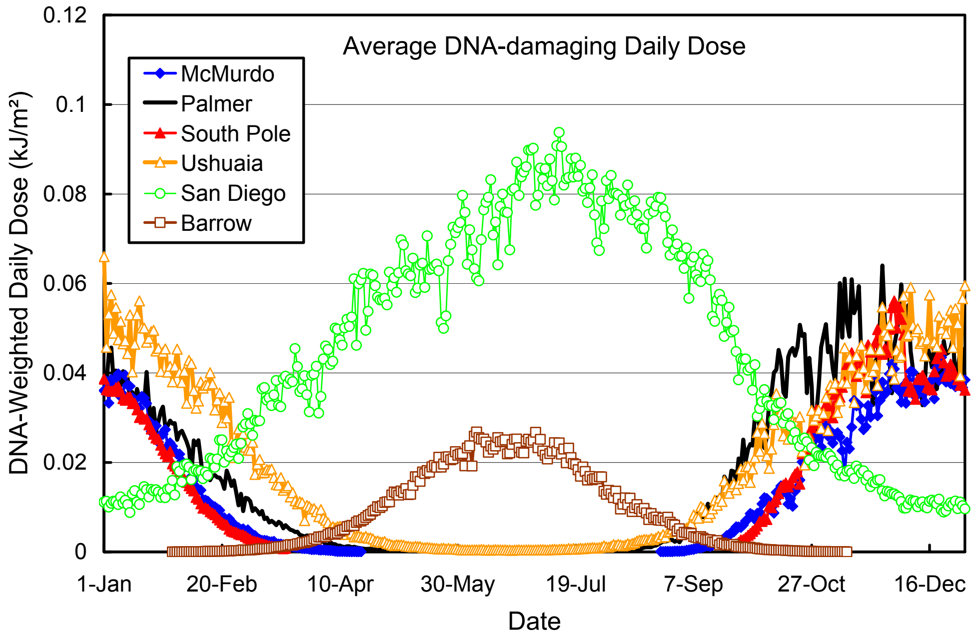

Figure 7.8.5. *Comparison of average DNA-weighted daily dose from all network sites. The data is based on average daily doses of the years 1991 – 2004.*

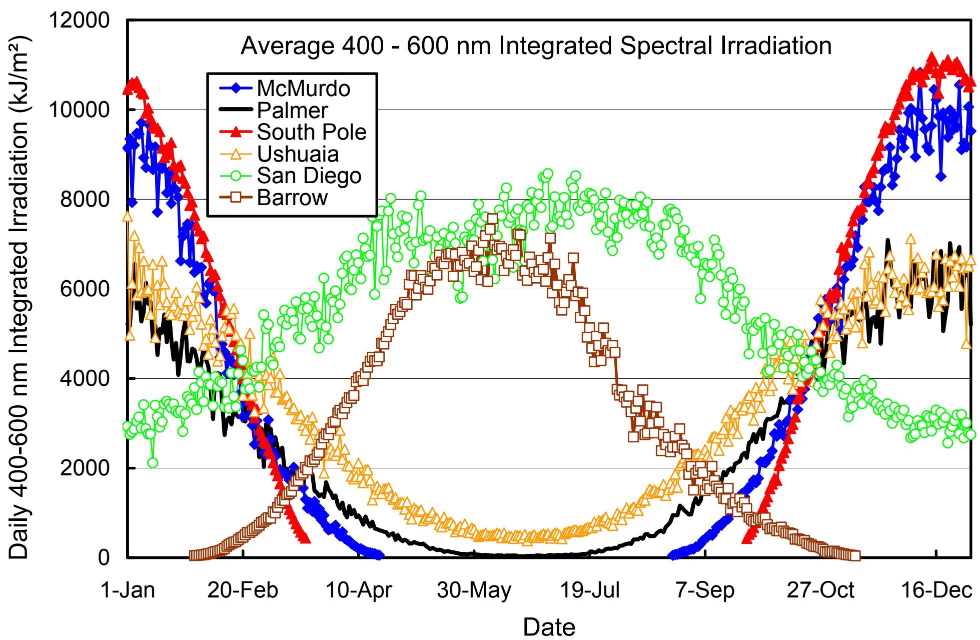

Figure 7.8.6. *Comparison of average daily irradiation of the 400-600 nm band from all network sites. The data is based on measurements from the years 1991 – 2004.*

7.9. Trends in UV

The assessment of trends in UV is affected by numerous factors, including the magnitude of the trend to be detected, and the variability and autocorrelation of the noise in the data (e.g. Weatherhead et al., 1998). Another important factor is the time span of available data.

Trends for the austral network sites have recently been calculated based on “Version 2” network data. For these calculations, data have been corrected for the instruments cosine-errors and residual wavelength shifts that affected early data of the monitoring network. These corrections also removed step-changes in data caused by system upgrades and modification to the data analysis software, which affected trend estimates published in previous Operations Reports. Please visit the Version 2 website at www.biospherical.com/NSF/Version2 for a detailed analysis of trends.

7.10. Factors Affecting UV Radiation

When UV radiation travels through the atmosphere it is partially absorbed by ozone, and scattered by air molecules, aerosol particles, and clouds. The cut-off of the solar spectrum in the UV-B is primarily a consequence of the sharp increase of the ozone absorption cross section toward shorter wavelengths. “Rayleigh-scattering” by oxygen and nitrogen molecules of the air is approximately proportional to (wavelength)$^{-4}$. This strong wavelength-dependence leads to a larger attenuation of incoming solar radiation at shorter wavelengths. In addition, ground-level irradiance becomes more diffuse as wavelength decreases.

In the following, the most important factors affecting UV radiation at the Earth’s surface are discussed. These are, in approximate order of decreasing importance: the extraterrestrial solar spectrum, solar zenith angle (SZA), clouds, total column ozone, ground albedo, atmospheric aerosols, and altitude.

Extraterrestrial solar spectrum

Temporal variations in the Sun’s radiative power in the UV-A and UV-B are smaller than 1%. However, due to the eccentricity of the Earth’s orbit around the Sun, solar irradiance at the top of the Earth’s atmosphere changes by ±3.4% during the annual cycle. The Sun is closest to the Earth on January 3 (Perihelion) and farthest away on July 4 (Aphelion). Extraterrestrial irradiance is therefore approximately 6.9% higher during the austral summer than during the boreal summer.

Solar zenith angle

The solar zenith angle (SZA) is defined as the angle between the zenith and the position of the Sun. It is a proxy for the average distance (or average path length) that photons travel from the top of the atmosphere until they reach the ground. Since most atmospheric absorption and scattering processes in the UV are more effective at shorter than at longer wavelengths, the change of ground-level UV irradiance as a function of SZA is more pronounced in the UV-B than in the UV-A. Moreover, the variation of UV with SZA is highly non-linear: a change of SZA by one degree at a small SZA leads to a smaller percentage change than at a large SZA.

Figure 7.10.1 shows the variation of global irradiance as a function of SZA at 300 and 400 nm. An additional line indicates the variation of spectral irradiance weighted with the erythema action spectrum. In this example, the largest change with SZA can be observed for spectral irradiance at 300 nm and 300 DU; irradiance at 400 nm shows the least change.

Variation in SZA cause most of the diurnal and annual variations in UV irradiance. The influence of the SZA is also responsible for most of the variation in UV with latitude. As a rule of thumb, erythemal irradiance increases by 5% per degree latitude, assuming all other parameters except SZA being constant. For mid-latitudes, erythemal radiation levels in summer are roughly one order of magnitude higher than in winter. As seen in the following example, the difference between winter and summer levels increases when moving to higher latitudes.

Noon-time SZA at summer solstice (June 21) is $L - 23.5°$, where L is latitude and 23.5° is the Earth's declination at solstice. Noon-time SZA at winter solstice (December 21) is $L + 23.5°$. For San Diego (33° latitude), noon-time SZA at summer and winter solstice are therefore 9.5° and 56.5°, respectively. Based on Figure 7.10.1, it can be estimated that summer erythemal values are higher by a factor of 4.4. Actual measurements from the NSF spectroradiometers show a difference of a factor of 4.2. This good agreement illustrates that SZA is the dominant parameter modulating UV at a site that is little affected by variations of ozone and albedo.

For Ushuaia (55° latitude), noontime SZA angles at summer and winter solstice are 31.5° and 78.5°, respectively. The difference between summer and winter erythemal irradiances estimated using Figure 7.10.1 is a factor of 26. Annually averaged measurements performed by the SUV-100 at Ushuaia indicate a difference of a factor of 24.5. However, the measured factor varies from year to year, because variations in the annual cycle of ozone are much more important at Ushuaia than in San Diego.

At latitudes beyond the polar circles (latitude 66.5°), the Sun is below the horizon at winter solstice, and a period of complete darkness (polar night) occurs. Differences between summer and winter UV levels are therefore extreme. Diurnal variations in SZA become generally smaller when moving closer to the poles.

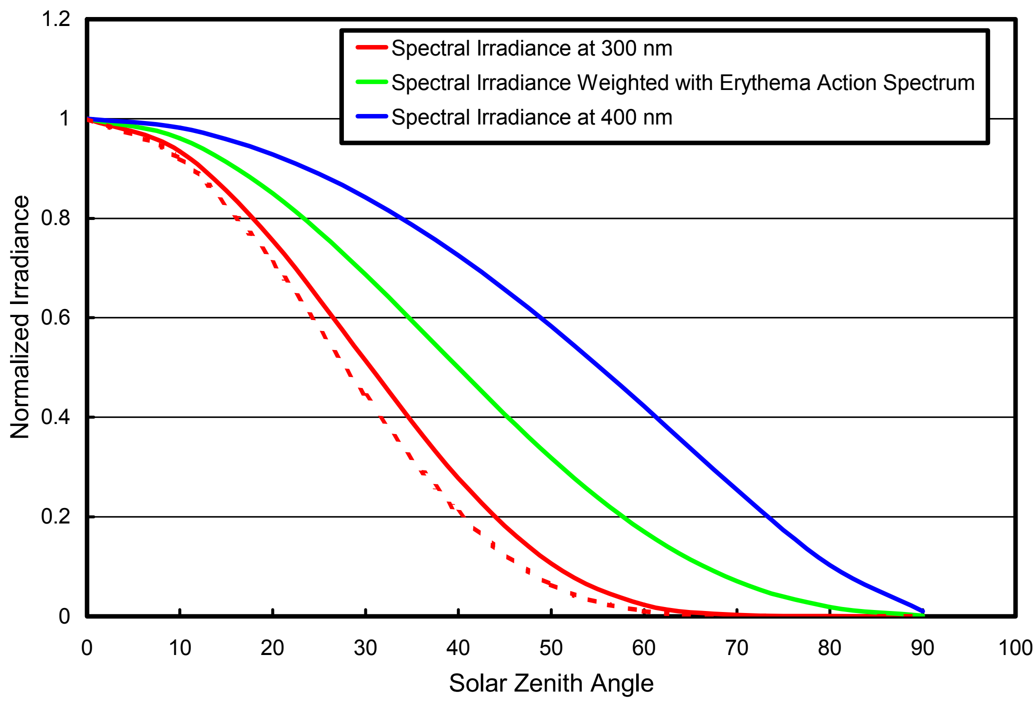

Figure 7.10.1: *Global spectral irradiance at 300 nm and 400 nm, and erythemally weighted irradiance as a function of SZA. The solid lines represent calculations with a total ozone column of 300 DU; the broken line was calculated with 400 DU.*

Clouds

The effect of clouds on UV is most difficult to parameterize because of their complex 3-dimentional character. The presence of clouds can lead to an increase or a decrease of ground-level UV. A uniform cloud layer will generally lead to a decrease of irradiance at the Earth's surface, because part of the radiation that is reflected upward by the cloud may escape into space. An increase may be observed if the disk of the Sun is not obstructed by clouds, and additional radiation is reflected from the side of a broken cloud field toward the ground (Mims and Frederick, 1994). Primarily due to the wavelength dependence of Rayleigh-scattering, UV radiation is more diffuse than visible radiation. The contribution to global irradiance from the direct beam of the Sun is therefore smaller in the UV. Clouds moving in front of the Sun lead consequently to a smaller attenuation of global irradiance in the UV than in the visible. Bais et al. (1993) and Blumthaler et al. (1994a) showed that with the solar disk clear of clouds, cloud amounts up to six octas produced little reduction in irradiance. Attenuation of UV irradiance as a function of cloud cover has been further parameterized by Thiel et al. (1997) and Josefsson and Landelius (2000). A correlation of the reduction of UV radiation by clouds versus the reduction of short-wave (300 – 3000 nm) solar irradiance as obtained by pyranometers has been investigated by Bordewijk et al. (1995), and Bodeker and McKenzie et al. (1996).

Seckmeyer et al. (1996) showed that optically thin clouds attenuate shorter UV wavelengths less than longer UV wavelengths, even though the reflectivity of clouds in the UV is approximately independent of wavelength. Kylling et al. (1997) explained this phenomenon with the wavelength dependence of Rayleigh-scattering: radiation that is reflected upward by the cloud is partly backscattered by air molecules situated above the cloud. Backscatter is more likely at shorter wavelength, and photons with shorter wavelengths therefore have a greater probability of reaching the top of the cloud a second time, and to traverse the cloud in a second attempt. In a similar study, based on SUV-100 data from Palmer Station and Ushuaia, Frederick and Erlick (1997) found that the wavelength dependence of cloud attenuation at those sites is caused by Rayleigh scattering occurring *between* the ground and the cloud base.

In contrast to optically thin clouds, optically thick clouds may lead to a stronger attenuation in the UV-B than in UV-A. Multiple scattering by water droplets in the cloud enhance the effective path length of photons, which in turn amplifies the effectiveness of ozone molecules inside the cloud in absorbing UV-B radiation (Mayer et al*.,* 1998).

Compared to all other factors, clouds are responsible for the highest short-term variability in UV. This introduces large errors if instantaneous surface radiation levels are to be estimated from satellite observations, which typically have a time-resolution of 3-24 hours only (Martin et al., 2000).

Total column ozone

Absorption by ozone is responsible for the cut-off of the solar spectrum in the UV-B. An increase in total column ozone leads to a decrease in UV levels, and this strong anti-correlation has been demonstrated repeatedly (WMO, 1999, and references therein). UV levels are also somewhat dependent on the vertical distribution of ozone in the atmosphere (i.e. the ozone profile). For small SZAs, a re-distribution of ozone from the stratosphere to the troposphere leads to a decrease in UV-B (Brühl and Crutzen, 1989). For very large SZAs, the re-distribution leads to an increase in UV-B. The effect has been further quantified by Lapeta et al. (2000) and Krzyscin (2000), who concluded that the range of variation of erythemal UV produced by the changes in the ozone profile is 5% at its maximum.

The relationship between change in total column ozone O_3 and change in biologically effective UV irradiance *E* can be quantified using "Radiation Amplification Factors" (*RAF*) (Booth and Madronich, 1994). For changes in ozone smaller than 10%, the RAF is defined as the relative fractional change in effective UV irradiance with fractional change in total column ozone:

$$RAF = -\left(\frac{\Delta E}{E}\right) / \left(\frac{\Delta O_3}{O}\right)$$

where ΔE and ΔO_3 are the respective changes of UV irradiance *E* and ozone O_3. For example, RAF=1 means, that a 1% decrease in ozone will lead to a 1% increase in effective UV. For changes in ozone exceeding 10% a power formulation of the RAF has to be used to describe the relationship between ozone column and UV irradiance more correctly:

$$RAF = \frac{\ln(E^* / E)}{\ln(O_3 / O_3^*)}$$

where E^* is effective UV irradiance corresponding to ozone column O_3^*, and E is effective UV irradiance corresponding to ozone column O_3. *RAF* coefficients typically vary between 0.1 and 4, depending on the biological effect. (See Madronich et al., 1991, and references therein for a comprehensive compilation of *RAFs* for various processes.)

The relationship between total column ozone and DNA-weighted irradiance is illustrated in Figure 7.10.2, which is based on measurements from all NSF network sites. The graph was constructed by comparing pairs of values with identical (or very similar) solar zenith angle, but different ozone column as measured by the TOMS satellite. Note that these datasets usually pair observations before and after summer solstice at the same solar angles to help isolate the impact of springtime ozone depletion. We determined RAF values for each site with a least square fitting routine. The results show that RAF values depend only slightly on site, and range between 2.02 and 2.30, indicating that a 1% change in ozone leads to a 2% – 2.3% change in DNA-weighted irradiance.

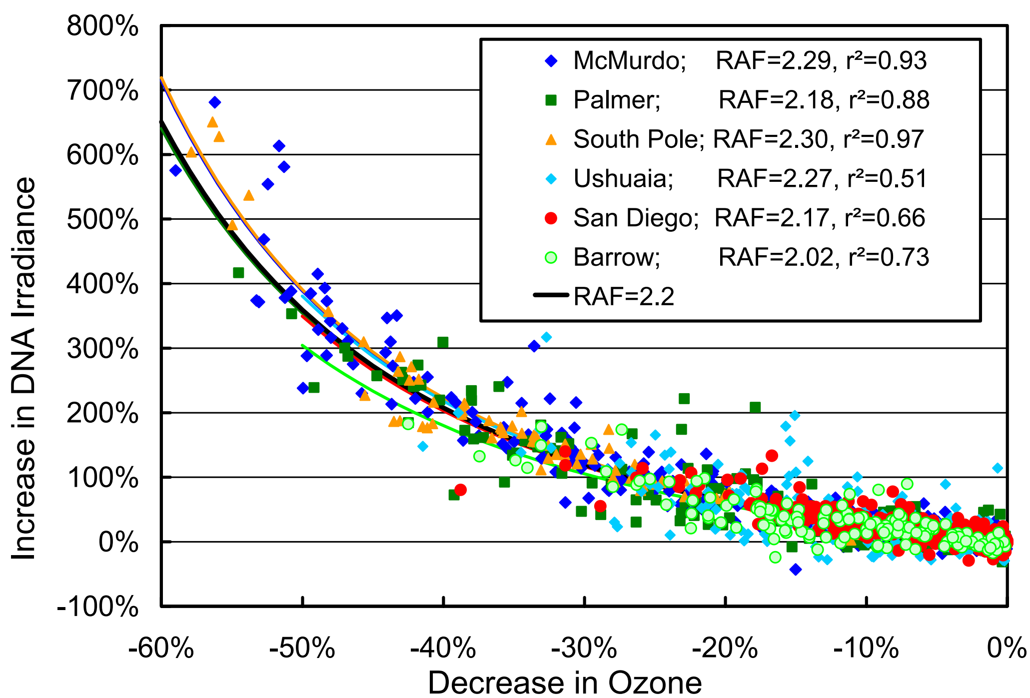

Figure 7.10.2. *Relationship between total column ozone and DNA-weighted irradiance. For each site, radiation amplification factors (RAF) were calculated by least square fit.*

The fact that the difference between sites is small suggests that RAF values depend only slightly on solar zenith angle and total column ozone. With the help of radiative transfer modeling, we studied this dependency in more detail. Model spectra calculated for solar zenith angles between 0° and 90°, and ozone values between 100 and 500 DU were convolved with the action spectra for erythema and DNA-damage. By correlating the biologically weighted UV values with ozone, RAF values were determined in dependence of SZA and total column ozone. The result is shown in Figure 7.10.3. RAF values for erythemal weighting vary between 1 and 1.2 for SZAs less than 70°. When the sun is lower, erythemal RAF values tend to decrease. This SZA and ozone dependence is of importance where prevailing SZAs are large, such as in the Arctic and Antarctic. In particular in spring, high total column ozone levels may lead to RAF factors that are significantly reduced compared to summer RAFs, or to typical RAFs for mid-latitudes. Changes in UV caused by trends in ozone (which are largest in spring) may therefore be mitigated by low Sun elevations.

For DNA-weighting, modeled RAF values range typically between 2 and 2.4, and are only lower for small ozone values and small SZAs. Since the DNA-damage action spectrum has a smaller contribution from the UV-A than the erythema action spectrum, it is more sensitive to ozone variations, leading to lower RAF values. RAF values determined theoretically agree well with RAF values determined with the data from the NSF network. For example for South Pole, RAF=2.3 was determined experimentally (Figure 7.10.2), and this values matches well the theoretically derived RAF values for high SZAs (Figure 7.10.3).

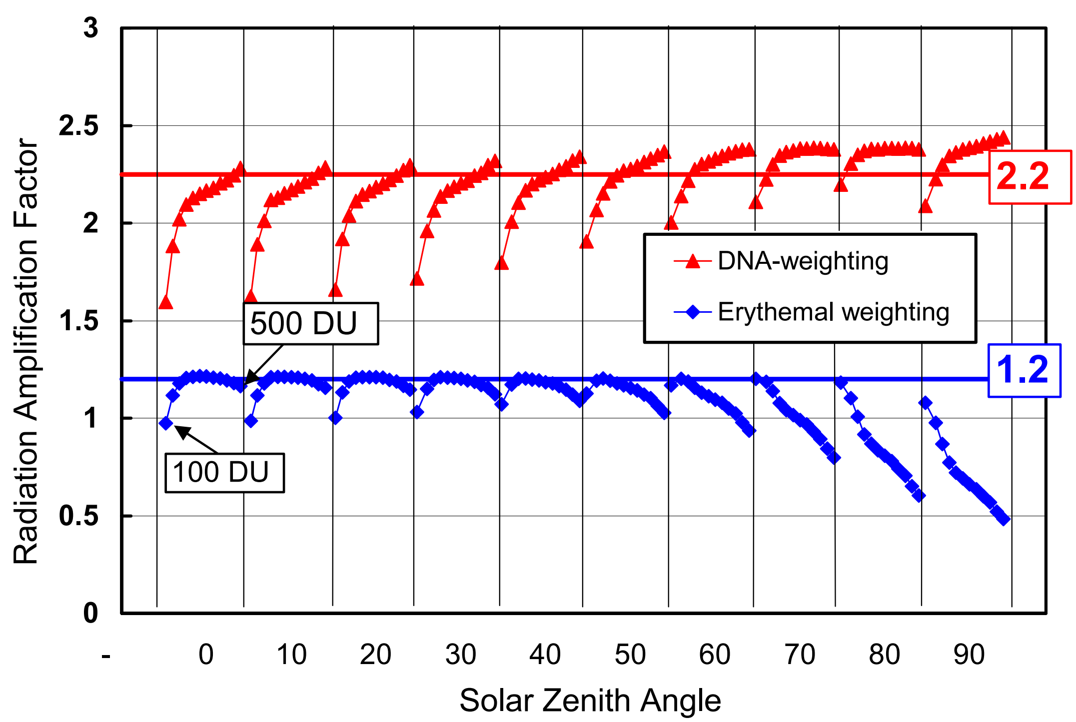

Figure 7.10.3. *Radiation amplification factor for DNA-weighted and erythemally weighted irradiance in dependence of SZA and total column ozone. The first value in each SZA-bin refers to a ozone column of 100 DU; the last value refers to 500 DU.*

Ground albedo

The reflectivity of the Earth's surface (albedo) leads to an increase in downwelling UV radiation as part of the radiation that is reflected upwards is scattered downwards by air molecules or clouds (Lenoble, 1998). For snow-free conditions, albedo in the UV is generally low and varies between roughly between 2% (grassland) and 10% (sand) (Blumthaler and Ambach, 1988). Snow-covered surfaces have an albedo between 50% and 98%. Dry, new snow has generally the highest albedo ranging between 90% and 98% (Grenfell et al., 1994). Open water has a UV albedo of about 5%; the albedo of sea ice is significantly higher and also depends on the amount of snow accumulation. Albedo is therefore an important factor for Arctic and Antarctic, where the ground is covered by snow during extensive periods of the year. Figure 7.10.4 shows the increase of erythemal global UV irradiance as a function of albedo for cloudless sky. The figure indicates that snow can increase erythemal irradiance by up to 60% compared to a case with negligible albedo. As scattering in the atmosphere may occur far away from the location of interest, the ground properties of a large area around the measurement site are relevant. The regional averaged albedo is often referred to as "effective albedo" (Kylling et al., 2000b; Gröbner et al., 2000), and may be defined as the albedo value used in a radiative transfer model that will lead to the best agreement of measurement and model. Three-dimensional radiative transfer models have shown that the area of significance (defined by increase of UV irradiance of more than 5%) can extend beyond 40 km (Ricchiazzi and Gautier, 1998; Degünther et al., 1998; Lenoble, 2000). Thus, even if there is fresh snow on the ground in the immediate vicinity of the measurement site, the presence of snow-free trees 5 km away may significantly reduce the effective albedo.

Albedo further mitigates the attenuation of UV radiation by clouds. Clouds are more effective in backscattering radiation than air molecules, as in the case of clear sky conditions. Multiple reflections between snow covered ground and clouds therefore lead to a significant increase of UV radiation at the Earth's surface compared to a snow-free situation (Kylling et al., 2000a).

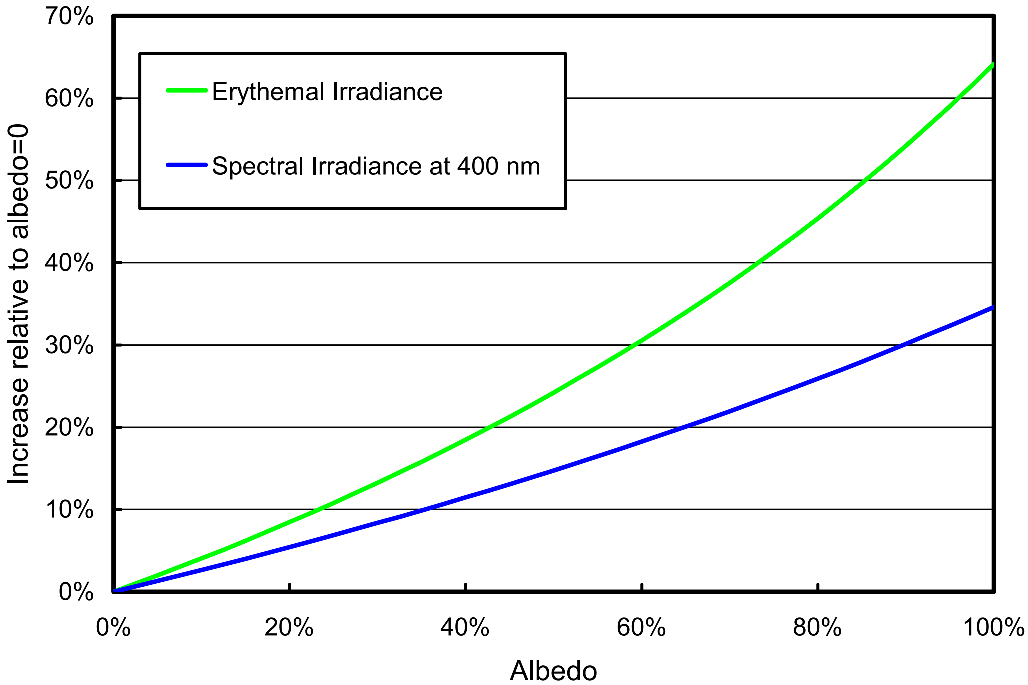

Figure 7.10.4. Increase of erythemal global irradiance and spectral global irradiance at 400 nm as a function of albedo. The graph is based on model calculations for SZA=60°, total column ozone = 300 DU, and clear sky. Spectral irradiance at 400 nm is less affected by Rayleigh scattering, and thus the enhancement by albedo is smaller than for erythemal irradiance.

Aerosols

Aerosols are solid or liquid particles suspended in air, and generally lead to a reduction of UV radiation. They are primarily located in the lower part of the troposphere. Some aerosols such as soot or mineral dust will partially absorb radiation, however, the primary reason of UV reduction is due to scattering. The ratio of the scattering cross-section to the total (i.e. scattering + absorption) cross-section is called the single scattering albedo, ω. For areas unaffected by industrial pollution, such as most of the Arctic and Antarctic, ω is typically larger than 0.95 (d'Almeida et al., 1991).

Unlike Rayleigh scattering, which results in an equal number of photons scattered in forward and backward direction, aerosols scatter photons primarily in the forward direction. The majority of photons are therefore scattered towards the Earth rather than towards space, and reach the ground as diffuse radiation.

The wavelength dependence of aerosol scattering can be approximated by Ångström's turbidity formula:

$$\tau_A(\lambda) = \beta\, \lambda^{-\alpha}$$

where $\tau_A(\lambda)$ is the aerosol optical depth at wavelength λ measured in µm. The coefficient α typically ranges between 0.5 and 2. A lower number indicates a larger average particle size. Note that α is significantly smaller than 4, which is the coefficient for Rayleigh scattering. This is to be expected since oxygen and nitrogen air molecules are smaller than aerosol particles.

The temporal variation of absorption by aerosols in relatively unpolluted areas such as the Antarctic is usually small compared to the influence of ozone and cloud cover. However, aerosols can have a significant impact on UV at Barrow, which is sometimes affected by a phenomenon known as "Arctic haze" (Bodhaine and Dutton, 1993). During such events, τ_A(500 nm) may increases from values smaller than 0.05 (background conditions) to 0.15-0.25, and α will change at the same time from values of 1.5 to values around 0.5 (CMDL, 2002). Model calculations indicate that a increase of τ_A(500 nm) from 0.05 to 0.2 leads approximately to a 5 % reduction in erythemal irradiance. These calculations assume α=1.3, ω=0.99, ozone column = 300 DU, low albedo, and no clouds. Reductions will be larger if aerosols are stronger absorbing (i.e. , ω<0.99). A quantitative assessment of the effects of aerosols on UV at Barrow is challenging due to the many parameters that affect aerosol attenuation and the difficulty of measuring these parameters accurately.

Altitude

UV radiation increases with altitude for several reasons. At higher locations, the atmosphere is "thinner," and therefore fewer particles exist that may absorb or scatter radiation. Higher locations also experience a reduced influence from tropospheric ozone or aerosols in the atmospheric boundary layer and the ground is more likely covered by snow. This leads to higher albedo, which in turn will increase UV. Clouds below a mountain summit have a similar effect as snow covered ground, and will therefore lead to an increase of radiation at the summit. In contrast, the same cloud may cause a decrease of radiation in a valley below the mountain. Since the variation of UV with altitude depends on various factors, which all have a different wavelength dependence, the increase of UV with height cannot be expressed by a simple relationship. Changes in erythemal irradiance with altitude reported in the literature therefore vary between 7%-25% per 1000 m altitude gain (Blumthaler et al., 1994b; Blumthaler et al., 1997; Gröbner et al., 2000; McKenzie et al., 2001).