2 Solar Radiation of the High Alps

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2.1 Introduction

Solar radiation at the Earth's surface varies by orders of magnitude that depend on actual local considerations. Therefore it is crucial to understand the influences of the various factors which determine the actual levels of solar radiation in order to estimate the effects on the whole biosphere. This is especially important for plants in the High Alps, as the levels of solar radiation are highest there due to a combination of several influencing factors.

While the spectrum of solar radiation gives the intensity at each individual wavelength, it is often sufficient to investigate the integral over a certain wavelength range. This is done e.g. for the integral between 280 and 3,000 nm, which is called 'total radiation', and also for the ranges of UVA (315–400 nm), UVB (280–315 nm) and UVC (100–280 nm). For all these integrations the weight of each individual wavelength is the same. This is in contrast to biological reactions, where radiation at different wavelengths usually has a different efficiency in triggering the reaction. Hence, in order to estimate the effects of radiation on the biosphere, one has to know the 'action spectrum' for each individual biological reaction under consideration. The action spectrum is usually a relative quantity in dependence on wavelength and normalised to unity at the wavelength with the maximum of the effect or at a standardised wavelength. The most common biological reaction in the UV wavelength range is the human erythema (McKinlay and Diffey 1987), which is often taken as a general measure for biological effects in the UV. Especially for reactions of UV on plants, a generalized action spectrum was determined (Caldwell et al. 1986), which is not very different from the erythema action spectrum. Therefore many of the findings relating to the variability of erythemally weighted irradiance can be interpreted in the same way for the generalized plant action spectrum. Furthermore, for plants the photosynthetically active radiation (PAR) is of high significance. This is the unweighted integral of radiation in the wavelength range of 400–700 nm. Due to the similarity to the wavelength range of visible radiation (defined by the spectral sensitivity of the human eye), the relations for visible radiation can also be interpreted to be relevant for PAR, and often the total radiation can also be taken as a good approximation for PAR.

The intensity of solar radiation can be measured with different geometries of the detector, corresponding to different applications. The most common type of measurement refers to a horizontal surface, which is called irradiance (intensity per unit area). The intensity of radiation falling on the horizontal surface at an angle to the vertical of the surface (zenith on the sky) is reduced according to the cosine of the zenith angle ('cosine law') due to the change of the projected area. This situation is valid for most types of surfaces, especially for human skin and for plants. For molecules in the atmosphere, which can be dissociated photochemically by radiation, the geometry is different, and in this case the 'actinic flux' (also called 'spherical irradiance') is the relevant quantity.

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Radiation from the whole sphere (4 pi) is received by the object or by the detector without any cosine-weighting, and radiation from all directions from the upper and lower hemisphere has the same weight.

2.2 The Sun

The solar spectrum outside of the Earth's atmosphere is close to the spectrum of a black body with a temperature of about 5,800 K. The spectral distribution follows the continuous spectral distribution of a thermal source with maximum irradiance at about 450 nm. However, the spectrum is significantly modified by absorption lines due to atomic absorption in the outer layers of the sun, the so-called Fraunhofer lines. These lines have a very high spectral structure far below 0.1 nm, so that the structure of spectral measurements depends very much on the spectral bandwidth of the instrument.

The total energy as the integral over all wavelengths from the ultraviolet through the visible to the infrared, which is received at the top of the Earth's atmosphere on a surface perpendicular to the radiation, corresponds to 1,367 Wm^{-2}. This value holds for the average distance between sun and Earth, it varies by $\pm 3.2\%$ during the year due to the elliptic path of the Earth around the sun, with the maximum value at beginning of January (minimal sun-earth distance) and the minimum value at beginning of July. In the UV wavelength range (100–400 nm) about 7% of the total energy is emitted, in the visible range (400–750 nm) about 46%, and in the infrared range (>750 nm) about 47%. However, the extraterrestrial spectrum is significantly modified by the processes in the atmosphere, as discussed in detail in the next section, leading especially to a smaller contribution of the UV range.

The intensity of the extraterrestrial solar radiation has some slight temporal variability on different time scales: 27-day solar rotation, 11-year cycle of sunspot activity and occasional solar flares. Mostly the UVC range is affected by up to a few percent (Lean 1987), whereas at longer wavelengths this variability can be neglected. Only an indirect effect may affect UVB levels at the Earth's surface due to effects on the strength of photochemical ozone production in the higher altitudes of the atmosphere.

2.3 The Atmosphere

The Earth's atmosphere mainly consists of nitrogen (78%) and oxygen (21%). Many more gases are present but only in very small amounts; however, they can still have a significant influence on the radiative properties (i.e. ozone, see discussion later). The molecules in the atmosphere scatter the light coming from the sun, which means that the direction of the propagation of photons is changed, but the wavelength (and the energy) is – for the purpose discussed here – unchanged. The probability of scattering increases with increasing density of the molecules (i.e. with increasing air pressure). The direction of photons scattered of molecules is mainly forward and backward (with equal amount) and less perpendicular to the propagation. Usually photons are scattered several times in the atmosphere ('multiple scattering'), which leads to a certain amount of 'diffuse' light. Of course, due to backscattering in the atmosphere, photons are also reflected back into space, and thus the total solar energy measured at the Earth's surface is less than the extraterrestrial one.

The probability for scattering of molecules (the so-called 'Rayleigh scattering') strongly increases with decreasing wavelength (λ) and can be approximated by a function proportional to λ^{-4} (Bodhaine et al. 1999). This means that the short blue wavelengths are much more efficiently scattered out of the direct beam of solar radiation than the red ones, and consequently blue dominates in the scattered light ('diffuse' radiation). Therefore the sky looks blue, if the air is clear and clean.

Besides scattering, molecules in the atmosphere can also absorb photons with specific wavelengths. In this case, the energy of the photons is transformed due to chemical reactions and finally transformed into heat, thus reducing the intensity of the radiation. As an example, ozone molecules in the atmosphere can absorb photons very efficiently in the UVC range and to a smaller extent also in the UVB range. This absorption by ozone is responsible for the fact that almost no radiation in the UVC can be observed at the Earth's surface, thus ozone acts as a protection shield against biologically very harmful radiation at these wavelengths.

In the Earth's atmosphere we also find 'aerosols', which are small solid particles of different sizes, shapes and chemical composition, which again can scatter and absorb solar radiation. The size of the aerosols is usually in the range of 0.05–10 µm. Being much larger than molecules, their scattering processes are different ('Mie-scattering'). In this case, the scattering probability depends only slightly on wavelength, as it is proportional to about $\lambda^{-1.3}$ (Angström 1964). Thus the scattered diffuse light due to aerosols is whiter (in contrast to the blue light scattered of molecules), producing a 'milky' sky in the event of large amounts of aerosols. Also, the direction of the scattered photons is different for aerosols compared to molecules; Mie-scattering has a very strong forward peak, so that the sky looks especially 'milky' in the surrounding of the sun. The absorption by aerosols is usually small, only a high concentration of soot (i.e. from forest fires) may lead to significant absorption.

Both together, scattering and absorption, are called 'extinction' and lead to a reduction of the intensity of the direct beam of solar radiation, and due to scattering to an increase in diffuse radiation. The extinction can be described with the 'Beer-Lambert-Law', where for the application relevant here, the scattering by molecules, absorption by ozone and extinction by aerosols are considered:

$$I = I_0 * \exp(-\tau_R(\lambda) * m_R - \tau_O(\lambda) * m_O - \tau_A(\lambda) * m_A)$$

'I' is the intensity at the Earth's surface, 'I_0' is the intensity of the incident radiation at the top of the atmosphere; 'τ' is the dimensionless 'vertical optical depth', which depends on the amount of molecules or aerosols (counted from the altitude of the observer to the top of the atmosphere) and which strongly depends on wavelength (λ); 'm' is the 'air mass', which characterises the pathlength of the photons through the atmosphere. If the sun is in the zenith, then $m = 1$, otherwise – for the assumption of a plane-parallel atmosphere – $m = 1/\cos(\text{sza})$, where sza is the 'solar zenith angle', which corresponds to (90° – solar elevation). For sza larger than about 85°, a correction of the atmosphere's spherical shape is necessary. The index 'R' marks the quantities for Rayleigh scattering of the molecules, 'O' for ozone absorption and 'A' for aerosol extinction.

As a consequence of the scattering processes in the atmosphere, the radiation field at the Earth's surface can be separated into the direct solar beam and the diffuse sky radiance. The angular distribution of the diffuse radiance is not homogeneous across the sky; it has a maximum around the sun, a minimum around 90° to the sun in the plane through the sun and the zenith, and especially for longer wavelengths it increases significantly towards the horizon (Blumthaler et al. 1996a). Both components together, direct solar beam and diffuse radiation, measured on a horizontal plane, are called 'global irradiance'. The share of diffuse radiation in global radiation is very variable (Blumthaler et al. 1994a). If the sun is completely hidden by clouds, all radiation is diffuse. Under cloudless conditions, the following dependencies can be summarised for the share of diffuse radiation in global radiation: it depends strongly (a) on wavelength with the highest values at the shortest wavelengths in the UVB, (b) on solar elevation with higher values at lower solar elevation (due to the longer pathlength) and (c) on the amount of aerosols with higher values at higher amounts. Furthermore, the share of diffuse radiation in global radiation is smaller at high altitude above sea level, because at higher altitudes the amount of scattering molecules (and aerosols) is lower due to lower air pressure, and thus scattering is less significant. As an example of these relations, measurements taken on a cloudless day at a high Alpine station (Blumthaler and Ambach 1991) indicated that for total irradiance and solar elevation above about 30° only about 10% of global irradiance is diffuse. These values can also be taken for PAR. In contrast, in the UVB-range generally about 50% of global irradiance is diffuse, and this value goes up to 90% and even 100% at very low solar elevations.

The separation of the radiation field into direct and diffuse components must also be considered if quantification of radiation on tilted surfaces is investigated. The contribution of the direct beam of solar radiation on a tilted surface can be calculated if solar elevation and horizontal angle of the position of the sun relative to the orientation of the tilted surface is known. The cosine of the angle between the sun and the vertical of the plane has to be considered in the calculation of the intensity of the direct beam (cosine law). The contribution of the diffuse radiation can be calculated straightforward under the simplified assumption that the diffuse radiance of the sky is homogeneous (Schauberger 1990). Otherwise this requires an extended calculation using a sophisticated radiative transfer model. As a tilted surface receives diffuse radiation from a part of the sky and also from a part of the ground, it is essential for the determination of the diffuse radiation falling on a tilted surface to know how much of the radiation is reflected from the ground (the so-called 'albedo'). This is of particular importance if the ground is covered by snow. The detailed discussion of the albedo follows in the next section.

An example of quantitative results was found in measurements on a vertical surface facing southward at the High Alpine Research Station Jungfraujoch in Switzerland (3,576 m). There the daily total erythemally weighted irradiance varied by a factor of 0.4–1.6 relative to a horizontal surface, depending on the season of the year (Blumthaler et al. 1996b). In winter time, when solar elevation is low, the vertical surface receives much more solar radiation compared to the horizontal one, whereas in summertime with high solar elevation the relation is the opposite. At a measurement campaign in Izana (2,376 m, Tenerife, Spain) the large range of variation of the irradiance ratio for vertical to horizontal surfaces was investigated as a function of wavelength and time of the day (Webb et al. 1999). At 500 nm an up to sixfold increase was observed when the solar elevation was low, whereas the increase was only about 20% at 300 nm under the same conditions. This is a consequence of the different relation between direct and diffuse irradiance, as discussed previously. Furthermore, when comparing the measurements from Izana and Jungfraujoch, the difference in ground albedo with snow at Jungfraujoch and snow-free terrain at Izania is also significant.

2.4 Variability of Solar Radiation Under Cloudless Conditions

The most important parameter determining the level of solar radiation under cloudless conditions is solar elevation. Furthermore, altitude above sea level, albedo (reflectivity) of the ground and amount and type of aerosols have a significant influence, and in addition the total amount of atmospheric ozone is of specific importance for the level of UVB radiation. Parameters that have only a minor effect on the level of solar radiation especially at higher altitudes, are the temperature profile in the atmosphere, the vertical distribution of the aerosols and the vertical profile of ozone. In the following sections, the effects of the main parameters are discussed in detail.

2.4.1 Effect of Solar Elevation

Solar elevation changes during the day, and the maximum value of solar elevation reached at solar noon depends on the season of the year and on the latitude of the observation site. When latitude, longitude, date and time are given, then solar elevation and azimuthal position of the sun can be calculated exactly. Of course, the higher the solar elevation the higher the level of solar radiation. However, this relation depends strongly on wavelength. UVB radiation is much more strongly absorbed at low solar elevations compared to radiation at longer wavelengths due to the longer pathlength within the ozone layer. Therefore, the diurnal course of UVB radiation is steeper compared to total radiation or PAR (Fig. 2.1).

For the same reason, the ratio between maximum daily values in summer relative to winter is also much higher for UVB radiation. Measurements at the High Alpine Research Station Jungfraujoch in Switzerland (3,576 m) have shown that the ratio of maximum daily totals in summer relative to winter is 18:1 for erythemally weighted UV irradiance and only 5:1 for total radiation and PAR (Blumthaler 1993).

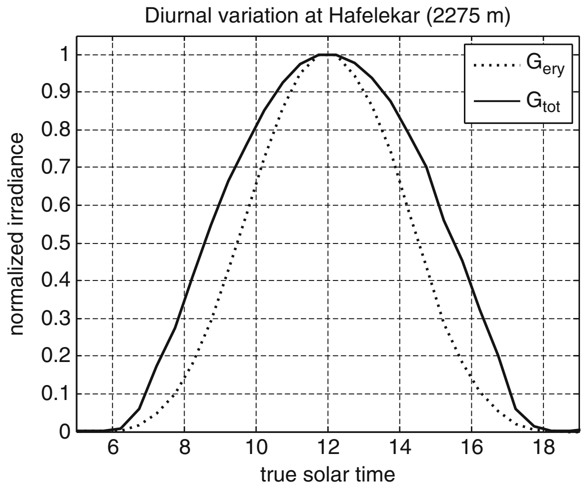

Fig. 2.1 Diurnal variation of erythemally weighted global irradiance (G_{ery}) and total global irradiance (G_{tot}) on a cloudless day at Hafelekar near Innsbruck (2,275 m above sea level). Both quantities are normalized to their maximum values at local noon

At high latitudes solar elevation is relatively low even in summertime. Therefore, the intensity of solar radiation is also relatively low, but the daily sum is significantly increased due to the longer length of the day.

2.4.2 Effect of Altitude

Generally global solar irradiance increases with increasing altitude above sea level. This increase is mainly due to a pronounced increase of direct irradiance, whereas for altitudes below about 3,000 m the diffuse irradiance is more or less constant (Blumthaler et al. 1997). The attenuation of direct irradiance due to extinction (following the Beer-Lambert-Law) becomes smaller, when the amount of scattering and absorbing molecules and aerosols becomes smaller. Furthermore, as the Beer-Lambert-Law depends strongly on wavelength, the increase of irradiance with altitude also depends strongly on wavelength. This increase with altitude is quantified with the ‘altitude effect’, which is defined as the increase of global irradiance for an increase in altitude of 1,000 m, relative to the lower site. Measurements of spectral irradiance at stations with different altitudes (Blumthaler et al. 1994b) show the dependence of the altitude effect on wavelength in the UV range (Fig. 2.2).

The strong increase towards the shorter wavelengths is a consequence of the smaller amount of atmospheric ozone at higher altitudes, although only about 9% of the total amount of atmospheric ozone is distributed in the troposphere. Additionally, the scattering on molecules (Rayleigh scattering) also increases strongly with decreasing wavelength. The figure also shows the relatively large range of variability of the altitude effect. This is a consequence of the strongly varying amount of aerosols and tropospheric ozone in the layer between the high and low altitude measurement stations. Under cloudless conditions, measurements in the Alps showed an altitude effect for UVB irradiance of 15–25% and of 10–15% for total irradiance (Blumthaler et al. 1992). In contrast, in the Andean mountains the altitude effect for UVB irradiance was only about 9% (Piazena 1996) or about 7% (Zarrati et al. 2003), because there the amount of aerosols and tropospheric ozone was very small. Furthermore, the altitude effect is additionally enhanced especially at shorter wavelengths if the ground is covered by snow at the mountain station and the ground is free of snow at the station in the valley (Gröbner et al. 2000).

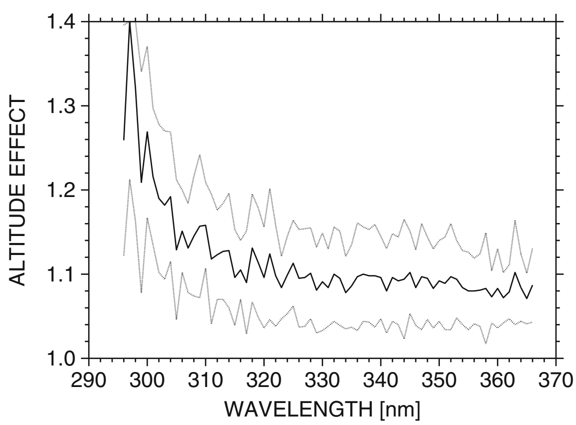

Fig. 2.2 Increase of spectral irradiance for an increase in altitude of 1,000 m, average (*solid line*) and range of variation (*dashed lines*) (From Blumthaler et al. 1994b)

As an example of the combined effect of solar elevation and altitude as discussed in 1.4.1 and 1.4.2, Fig. 2.3 compares the results of measurements of erythemally weighted UV irradiance (G_{ery}) and of total global irradiance (G_{tot}) at the High Alpine Research Station Jungfraujoch in Switzerland (3,576 m) and in Innsbruck (577 m).

The envelope of the seasonal course marks the cloud-free days with maximum values, whereas the other days are affected by clouds reducing the daily sum. Comparing the seasonal maximum values at Jungfraujoch and in Innsbruck shows the altitude effect, which is more pronounced in the shorter wavelength range. Comparing the seasonal course of total and erythemally weighted irradiance shows the different effects of solar elevation on the different wavelength ranges.

2.4.3 Effect of Albedo

The albedo of a surface is defined as the ratio of reflected irradiance to incoming irradiance. If the albedo is high, then the reflected irradiance is high, and consequently the diffuse irradiance is increased due to multiple reflections between the ground and the atmosphere. As a cloud layer will enhance these multiple reflections, the increase of diffuse irradiance due to albedo is highest under overcast conditions.

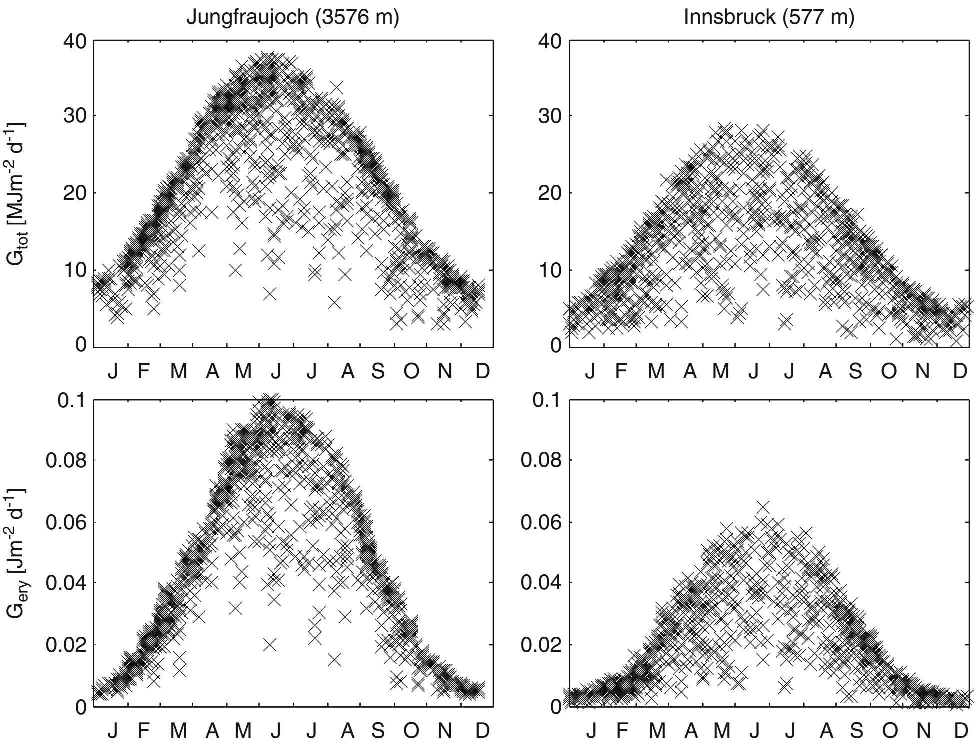

Fig. 2.3 Seasonal course of daily sums of total (G_{tot}) and erythemally weighted global irradiance (G_{ery}) at the High Alpine Research Station Jungfraujoch (Switzerland) and in Innsbruck (Austria)

The albedo depends on the type of surface, and also on wavelength for each surface. Spectral measurements of the albedo of various surfaces (Fig. 2.4) show a clear separation of albedo values: only snow-covered surfaces have a high albedo, which can exceed 90% in the case of fresh snow and be somewhat reduced if the snow is getting older and more polluted. For all other types of surfaces the albedo is relatively small and usually increases with increasing wavelength. The smallest albedo values were measured for green grassland, where the albedo was less than 1% in the UV range.

The increase of global irradiance due to a higher albedo can be quantified with an amplification factor, which gives the enhancement for a change in albedo by 10%. This factor generally increases with decreasing wavelengths, even if it is assumed that the albedo itself would be constant for all wavelengths, because the multiple reflections between atmosphere and ground are much more efficient for shorter wavelengths (Raleigh-scattering). However, in the UVB range, where ozone absorption is significant, the amplification decreases with decreasing wavelengths, because the longer pathlength of the photons due to multiple reflections will result in a more pronounced absorption by ozone. Under cloudless conditions the amplification of global irradiance due to a change of albedo by 10% is found to be about 1.03 at 300 nm, 1.035 at 320 nm and then decreasing to about 1.02 at 400 nm and 1.01 at 500 nm. As an example of the maximum effect of albedo, the enhancement of global irradiance for a change from green grassland to fresh snow is estimated to be about 30% at 320 nm and about 8% at 500 nm.

2.4.4 Effect of Aerosols

At higher altitude the amount of aerosols is usually relatively low, so that the effect of aerosols on global and diffuse irradiance is also relatively small. Aerosols mainly decrease the direct irradiance and increase the diffuse irradiance due to scattering. If the absorption of aerosols is high (e.g. for aerosols from biomass burning), then diffuse irradiance is less increased. Therefore, in many cases reduction by aerosols is

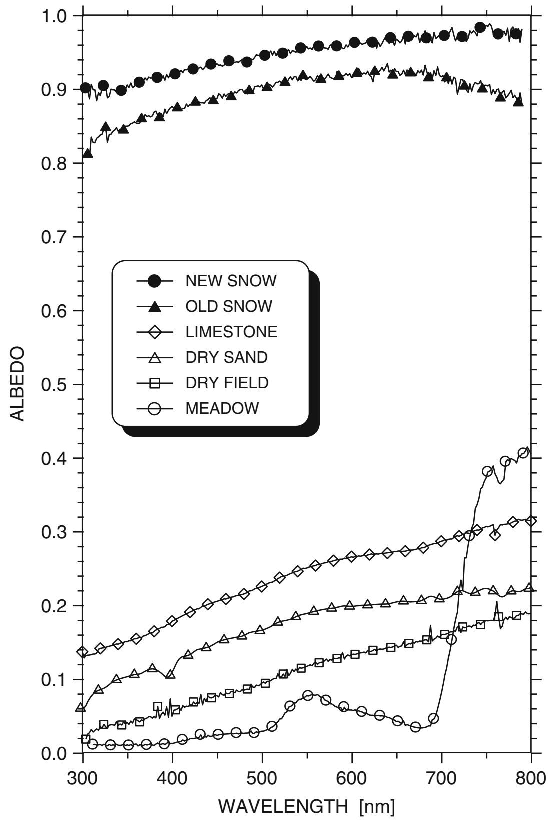

Fig. 2.4 Albedo of various types of surfaces as a function of wavelength (From Blumthaler 2007)

moderate (around 10–20%, significantly less at higher altitudes) for global irradiance, but occasionally it can be 30% or higher (Kylling et al. 1998). A special situation can occur if Saharan dust is transported far north and even up to the Alps. This type of aerosols is usually found at an altitude above the mountains and thus also affects the radiation in the High Alps, predominately scattering and only marginally absorbing, thus significantly increasing diffuse irradiance and only slightly reducing global irradiance.

2.4.5 Effect of Ozone

Ozone in the atmosphere mainly affects the UVB wavelength range due to its spectral absorption cross-section. For wavelengths higher than 330 nm its effect can be neglected, for shorter wavelengths it strongly increases with decreasing wavelength. This absorption by ozone in the atmosphere is the reason why almost no radiation with wavelengths below 285 nm can be measured at the earth's surface, although it is emitted by the sun.

At mid and high latitudes the total amount of ozone in the atmosphere varies strongly with time. Maximum values occur in springtime and minimum ones in autumn. Especially in springtime large day-to-day variability can occur with variations even larger than the seasonal ones, which results in significant variations of UVB radiation. To quantify the effect of ozone variations on UVB radiation, the so-called 'radiation amplification factor' is used, which gives the percentage increase of global irradiance for a decrease of ozone by 1%. This form is a simplification, which is in agreement with radiative transfer model calculations for ozone changes up to a few percent. The radiation amplification factor depends strongly on wavelength due to the wavelength dependency of the ozone absorption cross-section. For erythemally weighted irradiance the factor is about 1.1, for the generalized DNA damage it is 1.9, and for the generalized plant action spectrum it is about 1.8 (Madronich et al. 1998).

2.5 Variability of Solar Radiation Due to Clouds

In general clouds will of course attenuate solar radiation at the earth's surface, when a significant part of solar radiation is reflected back to space because of the high albedo of the top of clouds. Within a cloud, solar radiation is mainly scattered and usually only marginally absorbed. This scattering process within a cloud is almost independent of wavelength. The bottom of the cloud, as seen from the earth's surface, looks darker when the optical density of the cloud becomes higher, which depends on the density of the water droplets in the cloud. Very high clouds (Cirrus clouds, above about 7 km) usually consist of ice particles and always look relatively bright.

The effect of clouds on solar radiation at the earth's surface varies over a very large range, as does the density of the clouds and their distribution across the sky. In addition, it is of high significance whether clouds cover the sun itself or only the sky beside the sun. Under completely overcast conditions a reduction of global irradiance down to about 20–30% of the clear sky value as a rough average can be observed at low altitude (Josefsson and Landelius 2000), whereas at high altitude this reduction is somewhat less, the level being about 40–50% (Blumthaler et al. 1994a). This is caused by an average smaller thickness of clouds at higher altitudes. Although the scattering process itself within a cloud is almost independent of wavelength, the final attenuation of solar radiation becomes wavelength-dependent (Kylling et al. 1997). Solar UV radiation is attenuated up to 40% less than total radiation (Blumthaler et al. 1994a), which is mainly caused by the higher share of diffuse radiation at shorter wavelengths due to Raleigh-scattering.

However, under certain conditions clouds can also effect radiation enhancement at the earth's surface. If the sun itself is not covered by clouds, but big cumulus-type clouds are near to the sun, then reflections from the sides of the clouds may occur, which will enhance global irradiance at the surface. The degree of enhancement depends strongly on the local conditions, but for short time intervals an enhancement of total global irradiance of more than 20% can be observed (Cede et al. 2002). The enhancement is strongest for total radiation and about half the degree for UV radiation, as in the UV wavelength range the direct component (which is reflected from the sides of the clouds) contributes less to the global irradiance than in the total wavelength range.

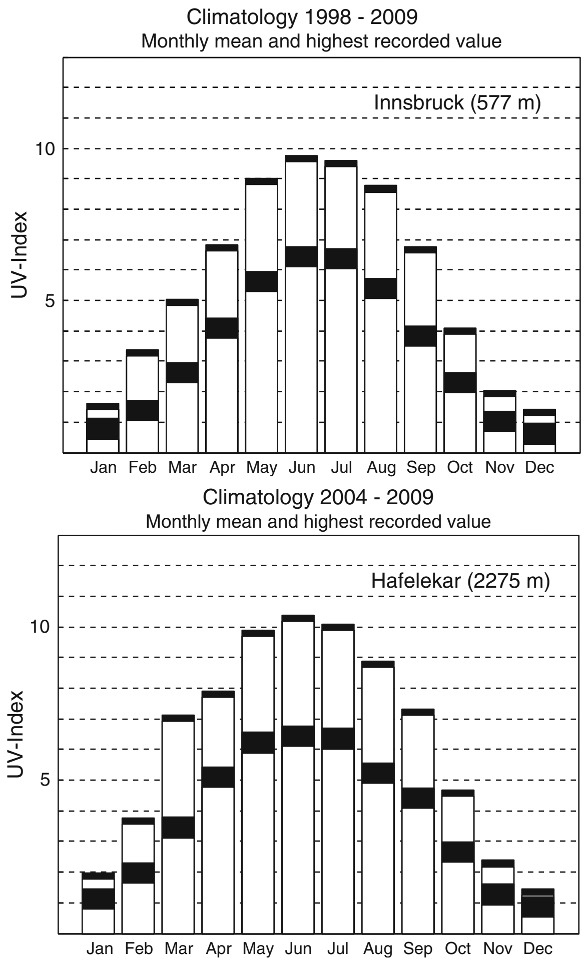

Fig. 2.5 Seasonal course of the monthly mean values of daily maximum erythemally weighted irradiance (*broad bars*) and the highest recorded value (*thin bar*) in Innsbruck, 577 m, Austria, (*top*) and at Hafelekar near Innsbruck, 2,275 m (*bottom*)

2.6 Climatological Aspects

In the previous sections the effects of individual factors on solar radiation were discussed; however, in reality it is always a combination of these factors that determines the actual level of solar radiation. In order to estimate the average radiation levels in the High Alps, longer time series of measurements are necessary. Although several decades of continuous measurements would be desirable to derive a complete climatology of solar radiation at a specific site, it is possible to discuss climatological aspects based on a shorter time series. For solar UV radiation the longest time series of measurements in Alpine regions are not much longer than one decade, but still they provide important characteristics. Figure 2.5 shows two examples taken from the Austrian UV monitoring network, which started to monitor the erythemally weighted UV radiation under different environmental conditions (urban/rural, low/high altitude) at several sites in Austria in 1998 and which includes 12 stations today. The raw data are collected every 10 min, and after conversion to absolute units they are published in near real time on the web site (www.uv-index.at). Following international recommendations, the erythemally weighted solar radiation is presented in units of the so-called 'UV-Index' (Global solar UV Index). This gives the erythemally weighted irradiance, expressed in W m^{-2}, multiplied by a scaling factor of 40, thus leading to values up to about 11 under the conditions of the Alpine environment. However, on a worldwide scale, UV index values up to 20 were observed in cities at high altitude in the Andeans.

In Fig. 2.5 the data for daily maximum values (expressed in the units of the UV index) are analysed as monthly averages (broad bars) for all available years of measurements. In addition, the thin bars indicate the maximum value of the UV index in each month, observed in any of the years of measurements. The top graph shows the results for Innsbruck (577 m), while the bottom graph shows the results for the nearby mountain station Hafelekar (2,275 m). As these data are average values, they include all weather conditions from cloudless to heavy rainfall or snowfall. It is quite surprising that the altitude effect as derived for cloudless conditions in 2.4.2 is almost invisible for the average values presented here. Thus on average the higher levels of radiation due to higher altitude can be masked by a higher average frequency of clouds at a mountain station. This might especially be the case for a station like Hafelekar, which is situated on a mountain ridge, where some convective clouds are frequently concentrated. Only the maximum values of the UV index are slightly higher at the mountain station (generally by less than 10%), which is again less than the altitude effect for cloudless conditions. The consequence of these analyses is that the average levels of radiation at higher altitudes depend very much on the local conditions of cloudiness, which can have a more significant influence on the intensity of solar radiation than the higher altitude. Furthermore, as clouds are the dominating parameter for the average values, the climatological results for erythemally weighted UV irradiance presented here can be generalized for PAR too.

Only for the months of March and April one can see significant differences between the measurement data for Innsbruck and Hafelekar as shown in Fig. 2.5. It is obvious that in these months the additional snow cover at the mountain station also leads to a significant increase for the average values presented here.

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