Doctoral Thesis

Solar erythemal ultraviolet radiation analysis of swiss measurements and modelling

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Solar Erythemal Ultraviolet Radiation.
Analysis of Swiss Measurements and Modelling

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Abstract

Small amounts of solar ultraviolet (UV) radiation are beneficial for people and essential in the production of vitamin D₃. Excessive UV exposure may lead to damage of the skin, eyes and the immune systems of humans and animals, as well as of terrestrial plants and aquatic systems. The discovery of the Antarctic ozone hole and the decrease of the ozone layer in the mid-latitudes have increased concern about the level of UV radiation.

When the solar extraterrestrial UV radiation enters the atmosphere, it interacts with the atmospheric constituents (air molecules, aerosol particles, droplets and ice particles). The variability in the different constituents as well as the position of the sun in the sky lead to the variability of radiation received on the ground.

The erythemal UV radiation is defined by the integral over the UV wavelengths of the irradiance multiplied by the erythemal action spectrum (carcinogenesis effectiveness).

In the present work, erythemal radiation measurements and radiative transfer calculations are analysed to estimate qualitative and quantitative impacts of changes in the atmospheric ozone content, aerosol particle loading, surface reflectance, altitude and cloud cover. The main data set is composed of the global, direct and diffuse erythemal UV irradiance measured at Davos (1610 m a.s.l.) and Payerne (490 m a.s.l.) since May 1995 (instrument: UV-Biometer Solar Light, model 501, 2-minute average values). Radiative transfer calculations are performed by using the model Tropospheric Ultraviolet-Visible (TUV).

The two most influencing parameters on clear-sky days are the solar zenith angle and the total ozone amount. Global erythemal irradiance is at Davos by factor 11.5 larger at 30° than at 70° solar zenith angle (300 DU ozone amount). For comparison, a decrease in total ozone from 360 to 240 DU leads to an increase by a factor of 1.4 at 30° solar zenith angle. The increase in erythemal radiation for a decrease in total ozone amount depends strongly on the zenith angle as well as on the total ozone amount.

Variation in the aerosol content influences the distribution of global irradiance into direct and diffuse irradiance. However, erythemal global irradiance varies only by maximal 5% between days with very low and very high aerosol...
If the ground is covered by snow on clear-sky days at Davos, erythemal irradiance increases by 15 to 25% due to multiple reflections between the surface and the atmosphere. This relative increase may reach 80% on overcast days.

The impact of clouds on radiation is highly variable (attenuation as well as enhancements). On overcast days at Davos with a snow free surface, erythemal UV radiation is reduced to a level which ranges between 8% (very thick cloud cover) and 70% (thin cloud layer) relative to clear sky. Under a broken cloud cover, the irradiance varies strongly and may be larger than the clear-sky value because of reflections on cloud sides.

The incidence of skin cancer strongly increased in the industrialized countries in the last decades, and cannot be explained by an increase in the UV radiation received on the ground. It has been shown to be due to increased exposure to sun light. The Swiss operational UV Index forecast model (collaboration with the Swiss Meteorological Institute) aims at raising public awareness regarding the risk of excessive exposure to UV radiation.
Résumé

Le rayonnement solaire ultraviolet (UV), en quantité limitée, est bénéfique aux êtres humains, et même essentiel pour la production de la vitamine D₃. Une surexposition peut toutefois altérer la peau, les yeux et le système immunitaire des humains et des animaux, ainsi que des plantes et des systèmes aquatiques. La découverte du trou d’ozone au-dessus de l’Antarctique et la diminution de la couche d’ozone dans les moyennes latitudes ont provoqué une prise de conscience concernant le rayonnement UV.

Lorsque le rayonnement solaire UV pénètre dans l’atmosphère, il interagit avec ses divers constituants (molécules d’air, particules d’aérosol, gouttelettes et cristaux de glace). La variabilité du rayonnement au sol est due à la variabilité des différents constituants de l’atmosphère et à la position du soleil dans le ciel.

Le rayonnement ultraviolet érythème est défini par l’intégrale de l’intensité du rayonnement multipliée par la sensibilité spectrale de la peau ("erythemal action spectrum", sensibilité associée à la carcinogenèse).


Les deux paramètres les plus importants qui déterminent le rayonnement érythème lorsqu’il n’y a pas de nuages sont l’angle au zénith du soleil et l’ozone. L’intensité érythème globale à Davos est 11,5 fois plus grande avec un angle au zénith de 30° qu’avec un angle au zenith de 70° (300 DU d’ozone). A titre de comparaison, une diminution de l’ozone de 360 à 240 DU avec un angle au zenith de 30° multipie le rayonnement par 1,4. L’augmentation du rayonnement érythème lors d’une diminution de l’ozone dépend fortement de l’angle au zénith ainsi que de la quantité initiale d’ozone.

La quantité de particules d’aérosol agit principalement sur la propor-
tion entre intensité directe et intensité diffuse du rayonnement. À Davos, l'intensité globale du rayonnement érythème varie au maximum de 5% seulement entre les jours avec un très faible et les jours avec un très grand trouble atmosphérique.

Lorsque la surface est recouverte de neige les jours sans nuages à Davos, l'intensité du rayonnement érythème augmente de 15 à 25% en raison des réflexions multiples entre le sol et l'atmosphère. Les jours couverts, l'augmentation due à la neige peut atteindre 80%.

L'influence des nuages sur le rayonnement est très variable (atténuation et intensification). Durant les jours couverts et sans neige, le rayonnement érythème est réduit à Davos à un niveau qui varie entre 8% (couche nuageuse très épaisse) et 70% (fine couche nuageuse) de la valeur par temps clair. Sous un ciel avec des nuages déchiquetés, le rayonnement varie fortement et peut même être plus grand que la valeur par temps clair s'il y a réflexion sur des faces de nuages.

Depuis quelques dizaines d'années, on observe une forte augmentation du nombre de cancers de la peau dans les pays industrialisés. Cette augmentation ne peut pas être expliquée par une augmentation du rayonnement UV au sol. Elle est due à l'augmentation de l'exposition au soleil. La prévision de l'index UV en Suisse (collaboration avec l'Institut Suisse de Météorologie) a pour but de rendre les gens concients des risques liés à une surexposition au rayonnement UV.
Zusammenfassung


Die erythemwirksame ultraviolette Strahlung ist definiert als das Integral der Bestrahlungsstärke multipliziert mit der spektralen Sensitivität der menschlichen Haut ("erythemal action spectrum", Kanzerogene Wirksamkeit).


Der solare Zenithwinkel und der Ozongehalt sind die Parameter, die die erythemwirksame UV-Strahlung bei wolkenlosem Himmel am stärksten beeinflussen. Die globale erythemwirksame Beststrahlungsstärke ist bei einem Zenithwinkel von 30° 11,5 mal größer als bei einem Zenithwinkel von 70° (300 DU Ozongehalt). Im Vergleich dazu nimmt die Strahlung um einen Faktor 1,4 zu, wenn der Ozongehalt bei einem Zenithwinkel von 30° von 360 auf 240 DU abnimmt. Die Zunahme der erythemwirksamen UV-Strahlung bei abnehmendem Gesamtozon hängt stark vom Zenithwinkel und vom Anfangsozongehalt ab.
Die Anzahl der Aerosolpartikel wirkt sich hauptsächlich auf das Verhältnis zwischen direkter und diffuser Bestrahlungsstärke aus. In Davos variiert die globale erythemwirksame UV-Strahlung zwischen sehr klaren Tagen und sehr trüben Tagen nur um maximal 5%.

Bei schneebedecktem Boden an wolkenlosen Tagen nimmt die erythemwirksame Bestrahlungsstärke um 15 bis 25% zu. Grund dafür sind Mehrfachreflexionen zwischen dem Boden und der Atmosphäre. An bedeckten Tagen kann die Strahlung aufgrund der Schneebedeckung um bis zu 80% erhöht werden.

Der Einfluss der Wolken auf der Strahlung ist stark variabel, wobei sowohl eine Abnahme als eine Zunahme möglich ist. An bedeckten Tagen ohne Schnee schwächt sich die erythemwirksame UV-Strahlung auf 70% (sehr dünne Wolkenschicht) bis auf 8% (sehr dicke Wolkenschicht). Die Strahlung variiert sehr stark unter einer aufgerissenen Wolkendecke. Wegen Reflexion an den Wolkenseiten kann die erythemwirksame UV-Strahlung größer sein als beim wolkenlosen Himmel.

Chapter 1

Introduction

Research in solar ultraviolet (UV) radiation finds its origin at mountain stations for the cure of tuberculosis (e.g. Davos and Leysin in Switzerland) (Suter, 1994). At the beginning of the 20th century, solar radiation, especially the shorter part of the spectrum (ultraviolet), was recognized as a therapy to treat tuberculosis (heliotherapy). The Physikalisch-Meteorologisches Observatorium Davos (PMOD) was founded in 1907 by Carl Dorno. Radiation at the shortest wavelengths reaching the earth’s surface (“Dornostrahlung”, nowadays UV-B) was investigated to find out why tuberculosis patients were better cured in high altitude stations than elsewhere. Inspired by cooperation with Carl Dorno, F. W. Paul Götz founded the Lichtklimatisches Observatorium (LKO) in 1921 to study the radiation climate of Arosa. Paul Götz performed measurements in the UV range and studied variations in total ozone content in collaboration with G. M. B. Dobson in England (see e.g. Götz 1929; Dütsch, 1992).

In the 60s, the cure of tuberculosis was based on pharmaceutical therapy and the ultraviolet radiation was recognized for its negative effect on the human skin. While small intensities of solar ultraviolet radiation are beneficial for people and essential in the production of vitamin D3, excessive UV exposure may lead to damage of the skin, eyes and the immune systems of humans and animals, as well as of terrestrial plants and aquatic systems.

The atmospheric ozone is a great absorber of the most damaging UV radiation (UV-B) and its depletion would lead to an increase of the UV-B radiation reaching the ground. Johnston (1971), Crutzen (1971) and Molina and Rowland (1974) demonstrated that anthropogenically released chlorine, containing compounds such as chlorofluorocarbons (CFCs), and bromine, containing volatile organic compounds such as halons, can cause stratospheric ozone depletion. In the mid 80s, a significant ozone depletion was indeed observed over Antarctica (the so-called ozone hole; Farman et al., 1985), and it could be shown that total ozone had decreased in the mid-latitudes as well, since the 70s (WMO, 1995; Staehelin et al., 1998b).
In response to this problem, an international consensus developed for the protection of the ozone layer. The Vienna Convention for the Protection of the ozone layer was signed in 1985, followed by the Montreal Protocol on Substances that Deplete the Ozone layer in 1987 as well as some amendments and adjustments. This international effort played a fundamental role in the decrease of the production and consumption of ozone depleting substances. At the present time, a stabilizing of the CFCs concentrations is observed in the troposphere (Montzka et al., 1996), while atmospheric bromine still continues to increase (Butler et al., 1998). Due to stratospheric cooling (greenhouse warming), however, the ozone column may continue to decrease about 15 years after the levels of stratospheric chlorine have started to decline (Shindell et al., 1998). Therefore, a stabilization of the ozone amount can be expected, but possibly with non-negligible time delay.

A depletion in stratospheric ozone is believed to lead to an increase of UV-B radiation reaching the ground. In turn, the increased intensities of shortwave UV radiation in the troposphere would then enhance photochemical reactions in the troposphere (see e.g. Gery, 1993), and also lead to damage of the ecosystem. Yet, the detection of expected trends is difficult because of the lack of reliable long series data and is complicated by the variability in the other components interacting with UV radiation in the atmosphere. Incidence of skin cancer strongly increased in the industrialized countries in the last decades (Slaper et al., 1996), and cannot be explained by an increase in UV radiation received on the ground. It has been shown to be due to increased exposure to sun light (vacation in low latitudes, excessive sunbathing, etc.; van der Leun et al., 1993).

Most of our knowledge about UV radiation comes from atmospheric radiative transfer models. These models combine measurements of the extraterrestrial solar radiation and atmospheric optics to calculate the amount of UV radiation reaching the surface at any time and location. The power of such models is that they allow detailed studies of the sensitivity of surface UV to numerous different conditions. Radiative transfer models are being continuously developed, with improvement of the numerical methods and modelling of the real atmosphere. Measurement of UV radiation, on the other hand, has been a very challenging issue because of very low intensities at ground level. Yet, in recent years, instruments and calibrations were greatly improved and reliable measurements are presently available, although they require regular and extensive care. Despite use of both efficient radiative transfer models and reliable measurements, however, the physical understanding of the behaviour of ultraviolet radiation in the atmosphere remains incomplete (e.g. uncertainties in the interaction with aerosol particles and mainly with clouds).

The purpose of the present PhD work is to improve our understanding of the interaction of UV radiation with real atmospheres. It is, however, linked to public health issues, since the main results concern erythemal UV radia-

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1Erythema: Abnormal redness of the skin caused by capillary congestion, following an
tion. Erythemal radiation measurements and radiative transfer calculations are analysed to estimate qualitative and quantitative impacts of changes in the atmospheric ozone content, aerosol particle loading, surface reflectance, altitude and cloud cover.

The main data set is composed of the global, direct and diffuse erythemal UV irradiance measured at Davos (1610 m.a.s.l.) and Payerne (490 m.a.s.l.) since May 1995 (instrument: Solar Light, model 501, 2-minute average values). The measurements are carried out in the framework of the Swiss Atmospheric Radiation Monitoring program (CHARM), a Swiss contribution to the Global Atmosphere Watch (GAW) program of the World Meteorological Organisation (WMO). CHARM aims at improving radiation measurements by developing the calibration and instrumental techniques, and studying the radiation budget in the Alps with its dependence on the altitude, season and cloud cover (Philipona et al., 1996). A detailed analysis is made possible by the existence of independent measurements of atmospheric ozone and atmospheric optical thickness, as well as of meteorological parameters at or close to the radiation stations. The simultaneous measurement of direct, diffuse and global erythemal irradiance allows the study of scattering processes, in addition to absorption processes.

Results based on measurements and radiative transfer calculations are applied in a short-term UV Index forecasting (erythemal UV level) for Switzerland (collaboration with the Swiss Meteorological Institute). This UV Index forecasting contribute at raising public awareness regarding the risk of excessive exposure to UV solar radiation.

Chapters 2 and 3 present the theoretical basis required for the understanding of the behaviour of the UV radiation (fundamentals and UV radiation in the atmosphere). Chapter 4 presents the data and radiative transfer model used for the analysis. Chapters 5 to 8 report results about the influence of atmospheric components and surface properties on erythemal UV radiation. Chapter 9 summarizes the Swiss UV Index short-term forecast procedure. Conclusion and outlook are presented in chapter 10. Some complements to the theory and results are gathered in four appendices.
Chapter 2

Fundamentals of radiation

This chapter introduces the concepts of the solar radiation theory required for the understanding of this report, as well as the notation. Further information about radiation theory, including the infrared part of the spectrum, can be found for example in Liou (1980), Kondratyev (1969), Chandrasekhar (1960) or Yanovitskij (1997).

2.1 Basic concepts

Electromagnetic radiation is described in term of its wavelength $\lambda$ and characterized by two related vectors: the electric ($\mathbf{E}$) and magnetic ($\mathbf{H}$) vectors.

Consider an element of area $dA$ and a differential solid angle $d\Omega$ oriented at an angle $\theta$ to the normal of $dA$ (Fig. 2.1). The differential amount of radiant energy $dQ_{\lambda}$ crossing $dA$ in a time interval $dt$ in the wavelength interval $(\lambda, \lambda + d\lambda)$ and in directions confined to $d\Omega$ is expressed in terms of the monochromatic radiance $L_{\lambda}$ [Jm$^{-2}$s$^{-1}$nm$^{-1}$sr$^{-1}$].

$$dQ_{\lambda} = L_{\lambda} \cos \theta \, d\Omega \, dA \, d\lambda \, dt$$ (2.1)

The monochromatic irradiance $E_{\lambda}$ [Wm$^{-2}$nm$^{-1}$] is defined by $L_{\lambda}$ integrated over the entire spherical solid angle: $E_{\lambda} = \int L_{\lambda} \cos \theta \, d\Omega$. In polar coordinates:

$$E_{\lambda} = \int_{0}^{2\pi} \int_{0}^{\pi/2} L_{\lambda}(\theta, \phi) \cos \theta \, \sin \theta \, d\theta \, d\phi$$ (2.2)

where $\phi$ is the azimuth, and $\theta$ the zenith angle. For isotropic radiation (radiance is independent of direction): $E_{\lambda} = \pi L_{\lambda}$.

We say light is polarized in a certain direction when the vibration of the electric vector $\mathbf{E}$ concentrates in that direction. The four Stokes parameters $(I, Q, U, V)$ are used to describe an elliptically polarized wave. These parameters are real numbers, function of the two complex components of $\mathbf{E}$.
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Figure 2.1: Illustration of pencil of radiation through an element of area $dA$ in directions confined to an element of solid angle $d\Omega$ and oriented at an angle $\theta$. Also shown is a solid angle and its representation in polar coordinates.

(parallel and perpendicular to a plane which includes the direction of propagation). The degree of polarization $P$ is defined as $P = \sqrt{Q^2 + U^2 + V^2}/I$. The light is linearly polarized if $V = 0$ and unpolarized if $Q = U = V = 0$.

2.2 Interaction of radiation with matter

When radiation interacts with a molecule or particle, a part of the incident energy is removed by absorption and scattering. We will assume that the interaction of radiation and matter does not give rise to a change of photon frequency.

2.2.1 Absorption

A molecule in the atmosphere can store various energies: (1) translational energy as a result of its motion in space, (2) rotational energy, (3) vibrational energy, as the atoms can vibrate about their equilibrium positions relative to the others, and (4) electronic energy, in case of the change of the energy state of the electrons. The last three energy types are quantized and take discrete values only. Absorption/emission of radiation takes place when the atoms or molecules undergo transitions from one energy state to another. This leads to complex absorption band systems associated with the molecular structure. Although the electronic energy variation is quantized, there is a broadening of the spectral absorption lines due to: the loss of energy or life time at the state, the perturbation due to collisions, and the Doppler effect resulting from the velocity distribution.
2.2. INTERACTION OF RADIATION WITH MATTER

The absorption cross section \( \sigma_a(\lambda) \) [cm\(^2\)] of a molecule for the wavelength \( \lambda \), analogous to a geometrical area, is the amount of energy removed by absorption from the original beam by that molecule.

### 2.2.2 Scattering

By scattering processes, molecules and particles in the path of electromagnetic waves continuously extract energy from the incident wave and reradiate that same energy in all directions.

The scattering cross section \( \sigma_s(\lambda) \) [cm\(^2\)] of a molecule or particle denotes the amount of energy removed by scattering from the original beam. Light emitted from the sun is unpolarized. After interacting with molecules and particles through scattering, the light becomes partially polarized.

### 2.2.3 Optical properties of an elementary volume

The optical properties of an elementary volume of medium are described by three quantities: the extinction coefficient \( k_\lambda \), the single scattering albedo \( \omega_0 \) and the phase function \( P \).

The extinction cross section \( \sigma_e(\lambda) \) is the sum of the scattering \( \sigma_s(\lambda) \) and absorption \( \sigma_a(\lambda) \) cross sections. The term mass extinction coefficient \( \beta_\lambda \) is used when the extinction cross section is in reference to unit mass [cm\(^2\)g\(^{-1}\)]. The (volume) extinction coefficient \( \beta_\lambda \) [cm\(^{-3}\)] refers to the extinction cross section multiplied by the particle number density \( N \) [cm\(^{-3}\)] or to the mass extinction cross section multiplied by the density \( \rho \) [g cm\(^{-3}\)]. The single scattering albedo \( \omega_0 \) is the ratio of the scattered to the extinct radiation. The phase function \( P \) describes the angular distribution of the scattered energy. It is a non-dimensional quantity with its integral normalized to \( 4\pi \).

\[
\int_0^{2\pi} \int_0^\pi P(\theta, \phi) \sin \theta \, d\theta \, d\phi = 4\pi \quad (2.3)
\]

or

\[
P(\theta, \phi) = 4\pi \frac{L(\theta, \phi)}{\int_0^{2\pi} \int_0^\pi L(\theta, \phi) \sin \theta \, d\theta \, d\phi} \quad (2.4)
\]

We also define the asymmetry factor \( g \) for azimuthal independent \( P \), as the average of the cosine weighted phase function (first moment of the scattering phase function).

\[
g = \frac{1}{2} \int_{-1}^{+1} P(\mu) \mu \, d\mu \quad (2.5)
\]

where \( \mu = \cos \theta \). The asymmetry factor ranges from -1 (100% backward scattering) to 1 (100% forward scattering).

The optical properties of particles and molecules are defined by the refractive index \( m \). It is composed of a real part \( m_r \) and an imaginary part \( m_i \) corresponding respectively to the scattering and absorption properties.
2.2.4 Reflection at a surface

When radiant energy is incident on a surface, it is partly absorbed, partly reflected, and partly transmitted. The reflected radiation is distributed among the reflection angles according to the bidirectional reflectance distribution function $R$ [sr$^{-1}$].

$$R_\lambda(\theta', \phi', \theta_0, \phi_0) = \frac{dL_\lambda(\theta', \phi')}{\mu_0 dL_\lambda(\theta_0, \phi_0)}$$

where $(\theta_0, \phi_0)$ is the incident angle, $\mu_0 = \cos \theta_0$, $(\theta', \phi')$ is the reflection angle, $\lambda$ the wavelength, $L_\lambda(\theta_0, \phi_0)$ is the incident radiance on a surface normal to the beam, and $L_\lambda(\theta', \phi')$ the reflected radiance in the direction $(\theta', \phi')$. Unless the surface has azimuthally dependent features, the dependence of $R$ on $\phi_0$ and $\phi'$ is reduced to a dependence on $\phi_0 - \phi'$.

The albedo $a_{\lambda}$ is the upward flux divided by the incoming radiation at a particular wavelength. It depends on the zenith angle of the incoming radiation.

$$a_{\lambda}(\theta_0) = \int_0^1 \int_0^{2\pi} R_\lambda(\theta', \phi', \theta_0) \mu' d\phi' d\mu'$$

The spectrally integrated albedo $\overline{a}_{\lambda}(\theta_0)$ is related to e.g. shortwave radiation, if the integral covers the shortwave range.

The two limiting characteristics of the reflective properties of a surface are called diffuse and specular. In the case of the perfectly diffuse surface (or Lambertian surface), the intensity of radiation leaving the surface is uniform in all directions (after reflection, the history of the incident radiation is completely obliterated). In the case of a perfectly specular surface, as a mirror, the incoming intensity contained within a solid angle $d\Omega_r$ and incident at an angle $\theta_i$ will be reflected within a solid angle $d\Omega_r = d\Omega_i$ and such that the reflected angle $\theta_r = \theta_i$.

2.3 Scattering of the radiation

2.3.1 Rayleigh scattering

Consider a small homogeneous, isotropic, spherical particle with radius much smaller than the wavelength of the incident radiation. The incident radiation produces a homogeneous electric field $E_0$ (applied field) which generates a dipole configuration on it. The electric field of the particle, caused by the electric dipole, modifies the applied field inside and near the particle. For unpolarized, incident radiation, the angular distribution of the light scattered by air is given by the Rayleigh phase function $P_{\text{Ray}}$.

$$P_{\text{Ray}}(\cos \Theta) = \frac{3}{4}(1 + \cos^2 \Theta)$$
2.3. SCATTERING OF THE RADIATION

where $\Theta$ represents the scattering angle (Fig. 2.2).

\[ \cos \Theta = \cos \theta \cos \theta' + \sin \theta \sin \theta' \cos (\phi' - \phi). \]

**Figure 2.2:** Illustration of the scattering angle $\Theta$ and its representation in polar coordinates. Note that $\cos \Theta = \cos \theta \cos \theta' + \sin \theta \sin \theta' \cos (\phi' - \phi)$.

The asymmetry of the air molecules produces a certain depolarization. A correction is applied using the depolarization factor $\delta$ (Eq. 2.9). Note that $\delta$ is a function of $m$, the nondimensional refractive index of molecules.

\[ P_{Ray}(\cos \Theta) = \frac{3}{4} \frac{2}{2 + \delta} \left( (1 + \delta) + (1 - \delta) \cos^2 \Theta \right) \quad (2.9) \]

### 2.3.2 Mie theory

If the particle dimensions are comparable with the wavelength of the incident light, the proper radiation field of the particle cannot be taken to be dipole. Depending on the relation between the dimensions of particles and the incident light, it is necessary to take account of fields of higher orders, such as quadropole or octopole, etc.

Mie (1908) has developed a theory assuming particles to be uniform spheres with a given complex refraction index $m$ and radius $r$. The Mie theory gives the extinction efficiency $Q_{ext} = \sigma_e(\lambda)/\pi r^2$, the scattering efficiency $Q_{sca} = \sigma_s(\lambda)/\pi r^2$, and the phase function $P$ of the particle in the form of series expansions. The absorption efficiency $Q_{abs} = Q_{ext} - Q_{sca}$, the single scattering albedo $\omega_0 = Q_{sca}/Q_{ext}$ as well as the asymmetry factor $g$ can be derived. Computer codes exist for the calculation of $Q_{ext}$ and $Q_{abs}$, as a function of the size parameter $x = 2\pi r/\lambda$ and $m$ (see e.g. Bohren and Huffman, 1983).

The most notable feature of Mie scattering is the "forward peak" for all but the smallest particles (Fig. 2.3). In the case of a tiny scatterer ($2\pi r << \lambda$), the scattering diagram would be symmetric, equal amounts scattered in backward and forward directions (Rayleigh scattering). If the particle size was increased, an asymmetry would begin to develop, implying more forward
than backward directed scattering. Holding the particle size constant, the forward peak is not greatly affected by an increase of the imaginary part of the refractive index $m_i$. The backward hemisphere is most affected. However, when the absorption becomes very large ($m_i > 1$) the backward-scattered radiation goes through a minimum and thereafter increases again. Further information about the Mie theory may be found in e.g. Twomey (1977).

![Scattering polar diagrams](image)

**Figure 2.3**: Scattering polar diagrams, for $\lambda = 0.320 \mu m$. Radius = $0.01 \mu m$ (left, $x = 0.19$), radius = $0.1 \mu m$ (middle, $x = 1.9$), and radius = $1 \mu m$ (right, $x = 19$). Calculations with the code of Bohren and Huffman (1983) with $m = 1.5 + 0i$.

### 2.3.3 Scattering by nonspherical particles

Scattering by a nonspherical particle depends on the directions of the incoming and outgoing radiation, and on the orientation of the particle with respect to the incoming beam. To include the effect of polarization, a $4 \times 4$ phase matrix $P(\theta)$ is required, which can be reduced to a $3 \times 3$ matrix as the ellipticity is very small (needs birefringent substances). The number of elements can be reduced in case of symmetries. Light scattering by nonspherical particles such as hexagonals crystal have been computed by e.g. Takano and Liou (1989a, 1995).

### 2.4 Radiative transfer in the atmosphere

The radiative transfer in the atmosphere is described by a general equation. Approximations of the equation were developed to calculate the transfer in specific cases.

#### 2.4.1 General equation of transfer and Beer-Bouguer-Lambert law

A pencil of radiation of wavelength $\lambda$ traversing a medium will be weakened by its interaction with matter with an extinction coefficient $k_{\lambda}$. The
intensity may also be augmented by emission of the material plus multiple scattering from all other directions into the pencil under consideration. The radiance \( L_\lambda \) becomes \( L_\lambda + dL_\lambda \) after traversing a thickness \( ds \) in the direction of its propagation. The relation between the incoming radiation \( L_\lambda \) and the variation of radiation \( dL_\lambda \) is given by the general equation of transfer.

\[
\frac{dL_\lambda}{k_\lambda \rho ds} = -L_\lambda + J_\lambda
\]  

(2.10)

where \( \rho \) is the density of the material, \( k_\lambda \) denote the mass extinction cross section for radiation of wavelength \( \lambda \), and \( J_\lambda = j_\lambda/k_\lambda \) is the source function (\( j_\lambda \) source function coefficient). The Beer-Bouguer-Lambert law (history in Middleton, 1964) is derived for a homogeneous medium, when both scattering and emission contributions to the pencil may be neglected (\( J_\lambda = 0 \)). It is given by

\[
L_\lambda(s_1) = L_\lambda(0) \exp(-k_\lambda u)
\]

(2.11)

where \( L_\lambda(0) \) is the incident radiation at \( s = 0 \), \( L_\lambda(s_1) \) is the emergent radiation at a distance \( s_1 \), and \( u = \int_0^{s_1} \rho ds \) is the path length.

The optical path length between two points \( s_1, s_2 \), also called optical depth \( \tau \), is defined by

\[
\tau_\lambda(s_1, s_2) = \int_{s_1}^{s_2} k_\lambda \rho ds
\]

(2.12)

where \( k_\lambda \) is the mass extinction cross section and \( \rho \) is the density. Note that the optical depth is often used as the vertical coordinate in radiative transfer equations (in place of the height) since \( d\tau_\lambda = k_\lambda \rho ds \).

The Beer-Bouguer-Lambert law depends only on the optical path length \( \tau_\lambda \) and is also applicable to the irradiance.

The monochromatic transmissivity \( T_\lambda = L_\lambda(s_1)/L_\lambda(0) \) of a given medium is defined as the ratio between the emergent and the incident intensities. The absorbivity \( A_\lambda \) represents the fractional part that is absorbed. We may also define the reflectivity \( R_\lambda \) as the ratio of the reflected (backscattered) intensity to the incident intensity. On the basis of the conservation of energy, we have \( T_\lambda + A_\lambda + R_\lambda = 1 \). The attenuation is defined as \( 1 - T_\lambda \).

### 2.4.2 Multiple scattering in plane-parallel atmosphere

In a realistic atmosphere, molecules and particles not only undergo single scattering but also multiple scattering (scattering more than once). Multiple scattering is an important process for the transfer of radiant energy in the atmosphere. We will assume that there are no internal sources, and no coherence between the radiation scattered from different particles.

Consider a plane-parallel atmosphere illuminated by flux radiation \( \pi L_0 \) emitted from the sun (see Fig. 2.4). Consider also \( \Omega \) a directional element of solid angle \( d\Omega \) that represent a pencil of radiation entering a layer (\( \tau \),
The differential diffuse intensity in $\Omega$ is reduced by single scattering and absorption. It may also be increased by multiple scattering, arising from the scattering of a pencil of radiation of solid angle $d\Omega'$ into the direction $\Omega$. In addition, the differential diffuse intensity in $\Omega$ also may be increased due to single scattering of the direct radiation whose direction is represented by $-\Omega_0$ (the minus sign denotes that the direct solar radiation is always downward). The amount of direct radiation at the layer $\tau$ is given by the Beer-Bouguer-Lambert law.

\[
\pi L_0 e^{\omega_0}\frac{d\omega_0}{d\Omega} \leq 0
\]

**Figure 2.4:** Scheme of the radiative transfer equation in a parallel layer.

Introducing the optical depth $\tau$ and single scattering albedo $\omega_0$, we obtain the radiative transfer equation in a plane-parallel layer.

\[
\mu \frac{dL(\tau; \Omega)}{d\tau} = L(\tau; \Omega) - \frac{\omega_0}{4\pi} \int_{4\pi} L(\tau; \Omega') P(\Omega'; \Omega') d\Omega' - \frac{\omega_0}{4\pi} \pi L_0 P(\Omega; -\Omega_0) e^{-\tau/\mu_0}
\]

(2.13)

Note that $\mu = \cos \theta$, $\mu_0 = \cos \theta_0$, $d\Omega = d\mu d\phi$, $\Omega = (\mu, \phi)$, and $\phi$ represents the azimuthal angle.

The equation (2.13) cannot be solved explicitly. The classical methods used to get a solution are: the method of principles of invariance which works with reflection and transmission functions, the adding method which uses a geometrical raytracing, the Monte Carlo method which models the history of individual photons, and the discrete-ordinate method presented in the next section. Note also that radiative transfer with polarization requires a matrix phase function instead of a scalar one.

A good summary of the classical methods, with advantages and disadvantages, is given in van de Hulst (1980) and Lenoble (1985). Note that the
information about the optical properties of the atmosphere at each height (profile) is needed to solve the radiative transfer equation.

2.4.3 The discrete-ordinate method

The discrete ordinate method for radiative transfer (DISORT) proceeds by expansion of the phase function and the radiance in spherical harmonics. The phase function is assumed to be azimuthal independent and its transformation from the scattering geometry into the geometry of plane-parallel atmosphere additive. Theorems for Legendre polynomials can be used to express the cosine of the scattering angle $\Theta$ as function of $\theta$, $\theta'$, $\phi$ and $\phi'$. Thus equation (2.13) splits up to $(K+1)$ independent equations (details e.g. in Liou, 1980).

\[
\frac{dL^k(\tau; \mu)}{d\tau} = L^k(\tau; \mu) - (1 + \delta_{0,k}) \frac{\omega_0}{4} \sum_{l=k}^{K} \hat{\omega}^k_l P^k_l(\mu) \int_{-1}^{1} P^k_l(\mu') L^k(\tau; \mu') d\mu' - \frac{\omega_0}{4\pi} \sum_{l=k}^{K} \hat{\omega}^k_l P^k_l(\mu) P^k_l(-\mu_0) \pi L_0 e^{-\tau/\mu_0}
\]

(2.14)

where $k = 0,..,K$; $\hat{\omega}^k_l$ are a set of constants, and $\delta_{0,k} = 1$ if $k = 0, 0$ otherwise. The Gauss quadrature (Eq. 2.15) is used to approximate the integral in Eq. 2.14 by placing on the interval $-1 \leq \mu \leq 1$ a set of $2n$ ($n \neq 0$) points $\mu_j$

\[
\int_{-1}^{1} f(\mu) d\mu \approx \sum_{j=-n}^{n} a_j f(\mu_j)
\]

(2.15)

where the $a_j$ are functions of the zeros of Legendre polynomials.

The value of $n$ determines the number of streams considered for the solution of the equation. The resulting system (2.14) with (2.15) is composed of $2n$ equations (one layer) and may be solved for $L^k$. The final solution is $L(\tau; \mu, \phi) = \sum_{k=0}^{K} L^k(\tau; \mu) \cos k(\phi_0 - \phi)$.

Stamnes et al. (1988) have presented a numerically stable algorithm for radiative transfer calculations in an atmosphere composed of plane-parallel layers. Spherical and pseudo-spherical improvements have been also implemented to take into account the sphericity of the atmosphere (Dahlback and Stamnes, 1991).

2.4.4 Two-stream methods

The two-stream methods are approximations of the DISORT method with the assumption that $n = 1$ (two streams). Two simultaneous equations are
derived for each layer, assuming the \( \mu \) dependence of \( L \) and approximating some integrals.

\[
\frac{dL}{d\tau} = \gamma_1 L^1 - \gamma_2 L^1 - \pi L_0 \omega_0 \gamma_3 \exp(-\tau/\mu_0) \quad (2.16)
\]

\[
\frac{dL}{d\tau} = \gamma_2 L^1 - \gamma_1 L^1 - \pi L_0 \omega_0 \gamma_4 \exp(-\tau/\mu_0) \quad (2.17)
\]

where

\[
L^1 = \int_0^1 \mu L(\tau, \mu) \, d\mu 
\]

\[
L^1 = \int_0^1 \mu L(\tau, -\mu) \, d\mu 
\]

\[
L(\tau, \mu) = \int_0^{2\pi} L(\tau; \mu, \phi) \, d\phi 
\]

The constants \( \gamma_i \) are functions of the asymmetry factor \( g \) and of the single scattering albedo \( \omega_0 \). Note that \( \gamma_3 + \gamma_4 = 1 \) (energy conservation).

The Eddington approximation makes the assumption \( L(\tau, \mu) = L_0(\tau) + L_1(\tau) \mu \). It was improved by Joseph et al. (1976) with the delta-Eddington approximation. The delta-Eddington approximation uses a delta function to approximate the peak in the forward direction. As a result, the parameters \( g' = g/(1 + g) \), \( \omega'_0 = (1 - g^2) \omega_0/(1 - g^2 \omega_0) \), and \( \tau' = (1 - g^2 \omega_0) \tau \) replace \( g \), \( \omega_0 \), and \( \tau \) in the radiation transfer equation. In general, we assume that the results of delta-methods are accurate to 10% or better. Larger errors may occur at large zenith angles or for large values of the quantity being considered.

A summary of two-stream approximations can be found in Meador and Weaver (1980), and improvements in the calculations procedure in Toon et al. (1989).

### 2.5 Extraterrestrial radiation

Most of the energy from the sun is in the form of electromagnetic radiation. The distribution of electromagnetic radiation emitted by the sun and incident on the top of the atmosphere is called the extraterrestrial solar spectrum (Fig. 2.5). The spectral distribution of the radian energy at the top of the atmosphere peaks in the visible range from about 400-700 nm. The wavelengths larger than 700 nm define the infrared range. The integrated average rate of incoming radiation over all wavelengths is called the solar constant (although varying slightly). It amounts to about 1366 W/m² at normal incidence at the top of the atmosphere for average Sun-Earth distance.

UV radiation represents only about 1% of the sun’s total radiative output. However, it initiates and controls the chemical, radiative, and dynamic processes which determine the state of the earth’s middle and upper atmosphere.
2.6. POSITION OF THE SUN AND RELATIVE AIR MASS

The UV range is divided into the UV-C (100 - 280 nm), UV-B (280 - 320 nm), and UV-A (320 - 400 nm) ranges.

The Sun-Earth distance varies during the year from minimum (January 3) to maximum (July 5). As a consequence, the intensity of the solar energy arriving at the top of the atmosphere varies by about ± 3.5%.

The extraterrestrial solar radiation for average Sun-Earth distance (1 AU) also shows a temporal variation. The variability is increasing towards shorter wavelengths and is associated with the 11-year solar cycle, the 27 days rotation period of the sun, the solar magnetic sector structure, and short-period phenomena. However, these variations are negligible for the wavelengths which reach the ground (larger than about 290 nm) (Lean, 1991). One can find further information on the sun irradiance variability and measurements in Fröhlich and Lean (1998).

2.6 Position of the sun and relative air mass

The position of the sun in the sky is a well-determined astronomical factor. The zenith angle \( \theta \) or the solar elevation \( \pi/2 - \theta \) determines the pathlength of the solar beam through the atmosphere. It is the most important parameter for temporal and geographical variations of solar UV radiation at the ground level. The calculation of the zenith angle \( \theta \) and azimuth angle \( \phi \) of the sun (Fig. 2.6) at a given day, time in day and locality is a function of the latitude, declination and hour angle. Approximate formula are presented \( e.g. \) in WMO (1986). See Appendix A for the operational procedure used at Lichtklimatisches Observatorium (LKO) Arosa.
For most scattering and absorption processes in the atmosphere which concern the direct solar beam it is necessary to know the total mass or optical path of atmosphere (or of relevant substance in the atmosphere) which the beam traverses on its way down to the ground. The relative air mass $m_T$ is the ratio between the vertical optical path and the actual slant optical path. The relative air mass equals to 1 for an overhead sun ($\theta = 0^\circ$). For a plane-parallel atmosphere it is given by $(\cos \theta)^{-1}$. 

**Figure 2.6**: Illustration of the zenith and azimuth angles of the sun.
Chapter 3

Ultraviolet radiation in the atmosphere

This chapter reviews the processes and atmospheric constituents which influence the UV radiation in the atmosphere. The optical properties as well as radiative transfer calculations (TUV code described in chapter 4) illustrate the relative effect of each process. Further information is presented in the following chapters in relation with the results.

When solar extraterrestrial UV radiation enters the atmosphere, it interacts with atmospheric constituents. Scattering occurs by the air molecules, the aerosol and cloud particles. Absorption occurs in some air molecules and in the aerosol particles. The variability in the different constituents, as well as of the position of the sun in the sky, leads to a strong variability of the intensity of radiation received on the ground.

The radiation from the solar disk is called direct radiation. The scattered radiation is called diffuse radiation. A portion of this diffuse radiation goes back to space (diffuse upward) and a portion reaches the ground (diffuse downward). The sum of the direct and diffuse downward (on a horizontal surface) is called global radiation.

3.1 Biologically active radiation

The UV-B radiation is responsible for most of the effects of sunlight on the human body. It is the main cause of sunburn and tanning, as well as of the formation of vitamin D₃ in the skin. It also influences the immune system. UV-B radiation does not penetrate far into the body; most of it is absorbed in the superficial tissue layers of 0.1 mm depth. However, primary reactions in the superficial layers have consequences throughout the body (van der Leun and Gruijl, 1993).

If reactions lead to damage, repair processes may come into play. The skin
has the ability to adapt. The epidermal layers become thicker, and melanin pigment is formed and dispersed throughout the epidermis. However, the repair systems may themselves be damaged by increased UV-B exposure. In general, sunburn can be avoided by sensible behaviour. The skin needs to adapt from winter conditions to the increased UV-B irradiance in summer. The avoidance of sunburn depends on going through this process carefully.

There are various types of skin cancer. One class is formed by the cutaneous melanomas, the cancer of the pigment cells. The other main types are basal cell carcinomas and squamous cell carcinomas, cancers of the epithelia cells (non-melanoma skin cancers). The incidence of cutaneous melanomas is lower than that of the non-melanoma cancers (factor 10). The mortality rate is, however, much higher. A primary melanoma could occur in a skin region seldom receiving any sunlight. However, observations showed that sunlight also plays a role (sudden exposure to high doses during childhood could be an important factor). The incidence of the non-melanoma skin cancer increased strongly during the last decades. They are clearly correlated to sunlight exposure. They occur mostly in light-skinned people, and then predominantly on skin areas most exposed to sunlight, such as face. The wavelengths responsible for carcinogenesis are determined by experimental observations.

![Figure 3.1: Erythemal action spectrum (solid curve) (CIE, 1987), generalized DNA damage spectrum normalized to unity at 300 nm (dotted curve) (Setlow, 1974), and generalized plant damage spectrum (dashed curve) (adapted from Cadwell et al., 1983).](image)

The erythemal action spectrum has been defined by the Commission Internationale de l'Eclairage (CIE), as a result of various experiments on the human skin (Mc Kinlay and Diffey, 1987; CIE, 1987). The erythemal action spectrum (Fig. 3.1) confirms that the carcinogenesis effectiveness of the UV-B is much higher than larger wavelengths. Similar experiments have been
conducted to determine the action spectrum of DNA and plants (Fig. 3.1).

The instantaneous biological dose rate $E_{\text{biol}}$ is calculated as:

$$E_{\text{biol}} = \int \lambda A_{\lambda} E_{\lambda} d\lambda$$

(3.1)

where $A_{\lambda}$ is the wavelength dependence of the spectral response function (action spectrum) for the biological process of interest, and $E_{\lambda}$ is the irradiance at wavelength $\lambda$. The erythemal UV irradiance is defined as the integration over the UV-B and UV-A wavelengths of the irradiance multiplied by the erythemal action spectrum (dose rate).

In general, it has been accepted that the UV-B portion produces the more deleterious biological effects. However, the UV-A radiation has been shown to produce the same biological effects, most likely through alternative mechanisms (Treina et al., 1996). Other experiments with mice suggested that UV-A radiation delays UV-B induced skin tumor development when administered in small daily doses (Bech-Thomsen et al., 1994). For further information about the effect of the UV radiation on the human beings (skin and eyes), plants, aquatic systems and animals see Tevini (1993). For information on the radiation in solaria see Moan and Johnsen (1994). Some information about the occurrence and increases in the number of skin cancer may be found in Slaper et al. (1996). The history of the research about the biologically effectiveness of the solar UV radiation can be found in Urbach (1987).

### 3.2 Attenuation by gas molecules

Absorption by ozone and Rayleigh scattering explain the main attenuation of the UV radiation for clear-sky conditions. As a consequence of these processes no radiation with wavelengths smaller than about 290 nm reaches the earth's surface.

#### 3.2.1 Absorption in gas molecules

Besides ozone ($O_3$), the solar UV radiation is absorbed by $O_2$, $N_2$, $NO_2$, $SO_2$, $O$, and $N$. The absorption due to the electronic transitions of molecular and atomic oxygen and nitrogen, and ozone occur chiefly in the ultraviolet region. Most of the UV radiation below about 250 nm is absorbed in the middle atmosphere by oxygen and nitrogen species (see Fig. 3.2).

Atomic $N$ and $O$ exhibit absorption spectrum from about 1 to 100 nm. The absorption of $N_2$ occurs at wavelengths shorter than 145 nm with complicated spectra. The absorption spectrum of $O_2$ ($O_2 + h\nu \rightarrow O + O^*$) begins at about 260 nm and continues down to shorter wavelengths. A portion of the atomic and molecular oxygen and nitrogen becomes ionized.
The absorption between 200 and 300 nm occurs primarily by ozone $O_3$. Ozone has a strong absorption between 180 and 340 nm (Hartley band) and a weaker band between 300 and 360 nm (Huggins band). The absorption in the Hartley band varies significantly with the temperature, with larger absorption cross sections for higher temperatures (Molina and Molina, 1986). Nitrogen dioxide ($NO_2$) and sulfur dioxide ($SO_2$) also absorb in the UV-B and UV-A spectra (Fig. 3.3). However, the concentrations of these species are much smaller than ozone (Fig. 3.4).

### 3.2.2 Scattering by gas molecules

An accurate estimate of the scattering by gas molecules is required in radiation transfer calculations. The scattering by gas molecules is described by the Rayleigh theory (see Liou, 1980 for the theory, and Howard, 1964, for the life of Lord Rayleigh). The scattered intensity is dependent on the wavelength to the fourth power, which explains the blue of the sky.

The Rayleigh scattering cross section and total volume scattering coefficients have been tabulated and formula are given for various model atmospheres. They give results as a function of the wavelength, with incorporation of the variation of the depolarization factor with wavelength. Details can be found in Fröhlich and Shaw (1980) and Bucholtz (1995).

The Rayleigh scattering cross section $\sigma_r(\lambda)$ equals to $5.59 \times 10^{-26}$ cm$^2$ at 300 nm and $1.67 \times 10^{-26}$ cm$^2$ at 400 nm (WMO, 1985b). The volume-scattering coefficient $\beta_1(z) = N(z) \sigma_r(\lambda)$, where $N(z)$ is the molecular number density at altitude $z$ at given pressure and temperature, is presented for 300 and 400 nm in Fig. 3.5.
3.3 Absorption by atmospheric ozone

The UV-B radiation is strongly influenced by the atmospheric ozone content. Thus, the influence of ozone has to be carefully treated to explain the radiance of the erythemal UV radiation in the atmosphere and on ground.

The atmospheric ozone is distributed in the troposphere and in the stratosphere, with a maximum of partial pressure at about 20 - 23 km in the mid-latitudes (the so-called ozone layer). The column density of ozone or total ozone amount is expressed as an equivalent layer thickness in centimeters, at standard temperature and pressure. It is normally given in Dobson units (DU) (100 DU = 1 mm thickness). One can find a good summary of the history of the ozone research in Walshaw (1989) and Dutsch (1992).

3.3.1 Spatial and temporal variability of the atmospheric ozone

The distribution of the atmospheric ozone depends on photochemistry and atmospheric transport. Both are functions of latitude, altitude and season. Ozone is produced mainly over the tropics at 25-35 km altitude and then transported to the middle and high latitudes of the winter hemisphere. Called Dobson-Brewer circulation, this process is responsible for the seasonal variation, with maxima in spring and minima in fall (see Fig. 3.6).

Advection (vertical and horizontal), ascent and descent of air in the lower stratosphere and around the tropopause are mainly responsible for the spa-
Figure 3.4: Typical profiles of ozone (solid curve), molecular oxygen (dotted line) and nitrogen dioxide (dashed curve) [cm\(^{-3}\)]. U.S. standard atmosphere (Anderson et al., 1986).

Ozone "mini-holes" are observed in spring in the mid-latitudes. They represent regions of low total ozone that persist for several days. These events are usually associated with breaking Rossby waves in the upper troposphere. These breakings allow poleward injections of ozone-poor, low potential vorticity tropospheric air into the lower stratosphere (McCormack and Hood, 1997). In spring, ozone-poor air masses (ozone depleted by catalytic cycles, see below) and ozone-rich air masses are present in the arctic regions. The transport of these air masses downward to the mid-latitudes also leads to large variations in the total ozone content and profile.

The chemistry of the ozone layer was first investigated by Chapman (1930), who proposed a chemical scheme involving oxygen compounds (see Eq. 3.2). In this scheme, the production of ozone is balanced by photolysis and subsequent reaction of O with \(O_3\).

\[
\begin{align*}
O_2 + h\nu & \rightarrow O + O \quad (\lambda < 242 \text{ nm}) \\
O_2 + O + M & \rightarrow O_3 + M \\
O_3 + h\nu & \rightarrow O_2 + O \quad (\lambda \approx 200 - 320 \text{ nm}) \\
O + O_3 & \rightarrow 2O_2
\end{align*}
\]
3.3. ABSORPTION BY ATMOSPHERIC OZONE

Several ozone destruction cycles involving NO\textsubscript{x}, ClO\textsubscript{x}, BrO\textsubscript{x} and HO\textsubscript{x} have been then identified. The net effect of each cycle is 2O\textsubscript{3} + h\nu \rightarrow 3O\textsubscript{2}. Some cycles depend on the formation of the O radical and apply to the middle and higher stratosphere. Other cycles work at lower altitudes. The individual cycles are coupled with each other so that their effects on ozone do not simply add. The formation of reservoir gases contributes to the stability of stratospheric chemical equilibrium by suppressing catalytic ozone destruction. See the summary of the catalytic ozone destruction and coupling of cycles due to the formation of reservoir gases (ClONO\textsubscript{2}, HCl, HNO\textsubscript{3}) in Peter (1994, Table 1).

The Antarctic ozone hole is created by the conjunction of meteorological processes (polar vortex) and photochemistry. The cycle involving ClO\textsubscript{x} components together with very low temperatures is believed to be mainly responsible for the ozone depletion in the polar spring. Similar reactions with BrO\textsubscript{x} components may also be involved (with a less pronounced temperature dependence). Heterogeneous surface reactions on stratospheric cloud particles (formation at very low temperatures) add to the gas-to-gas chemistry in the activation of chlorine. Details about the ozone chemistry in the stratosphere and the Antarctic ozone hole may be found e.g. in Peter (1994) or Piechowski et al. (1994).

The observed depletion in the mid-latitudes (McPeters et al., 1996, Stachelin et al., 1998b) cannot be explained using gas-phase chemistry models. Thus, heterogeneous reactions might be responsible for the slow ozone decay. The ozone loss can be imported from further north or produced in situ by reactions with the background aerosols. If sufficient active chlorine become available, the ClO\textsubscript{x} cycle comes into play and in particular the
The ozone chemistry in the troposphere is related to the air pollution. Emissions of nitrogen oxides (NOx), volatile organic compounds (VOCs) and carbon monoxide (CO) lead to a production of ozone and other photooxidants in the presence of solar radiation (see details e.g. in Finlayson-Pitts and Pitts, 1986). The ozone concentrations near the ground are strongly related to the meteorological conditions and transports in the boundary layer. Large tropospheric ozone concentrations on sunny summer days (see Fig. 3.8) may cause damage to the vegetation, as well as problems for the human respiratory system.

3.3.2 Absorption of the UV radiation

The variation of the ozone concentration has a strong influence on the UV-B radiation reaching the earth’s surface (Fig. 3.9). Note, however, that this effect remains relatively small in comparison with the zenith angle influence. The erythemal UV radiation, which contains UV-B (absorbed in ozone) and UV-A (not absorbed in ozone) radiation, is strongly attenuated by the atmospheric ozone because of its high sensitivity in the UV-B range (Fig. 3.10).

The profile of the ozone also plays a role in the radiative transfer. The tropospheric ozone is assumed to be more efficient than the stratospheric ozone.
3.4 Attenuation by aerosol particles

Propagation of radiation through the atmosphere is affected by absorption and scattering by particulates (e.g. haze, dust, fog, and clouds) suspended in the air. Scattering and absorption by haze or aerosol particles becomes a dominant factor in the boundary layer.

3.4.1 Tropospheric aerosols

Aerosols in the boundary layer have the greatest variability. These aerosols consist of a variety of natural and man-made chemical compounds, inorganic as well as organic. Particles are transported from their source regions into the atmosphere, or may be formed within the atmosphere by gas to particle conversion through chemical or photochemical processes (Shettle, 1989). The aerosol particles condition is highly dependent on the meteorological condition (rain, inversion, stability). A variation in the relative humidity means also a change in the composition and affect the refractive index of the aerosol particle (see e.g. d’Almeida et al., 1991). Detailed description of the aerosol particle formation and scavenging can be found in Pruppacher and Klett (1997), or Twomey (1977).
3.4.2 Stratospheric aerosols

The aerosol particles normally present in the lower stratosphere (up to about 30 km) are largely sulfuric acid solution droplets (Shettle, 1989). They are produced through photochemical reactions involving carbonyl sulfide (COS), sulfur dioxide (SO₂), and other gases. The background stratospheric aerosols are perturbed by major volcanic eruptions which can inject signif-
3.4. ATTENUATION BY AEROSOL PARTICLES

Figure 3.9: Spectra of the global radiation [W m\(^{-2}\)] for 250 DU and 25° zenith angle (solid curve), 400 DU and 25° zenith angle (dotted curve), 250 DU and 70° zenith angle (dashed curve), and 400 DU and 70° zenith angle (dotted-dashed curve). TUV calculations (sea level, Elterman (1968) aerosol profile, low surface albedo).

Significant amounts of SO\(_2\) and volcanic ash into the stratosphere. The larger volcanic ash particles will settle out in a short period of time. However, the sulfuric acid aerosols produced from the SO\(_2\) will persist for several years, with a 1/e loss time of about 12 months. The background stratospheric aerosol particles have a negligible effect on the surface UV radiation. The influence may increase after a volcanic eruption.

Michelangeli et al. (1989, 1992) noted that the radiation reaching the surface can be increased by scattering by stratospheric aerosol. They attributed this increase to “photon trapping”, which refers to the amplification of downwelling irradiance. “Photon trapping” occurs when the multiple scattering of photons caused them to pass a given level more than once. Davies (1993) noted that this enhancement can be explained by single-scattering theory when the atmosphere below the aerosols is highly absorbing and for large zenith angles.

3.4.3 Redistribution of the direct in diffuse radiation

The atmospheric particles vary greatly in the concentration, size, composition, and consequently in their effects on radiation. The influence of aerosols only slightly depends on wavelength, with a greater effect at lower wavelengths. The UV range is seriously affected by aerosols. However, ozone absorption and molecular scattering are much more important in most cases.

The aerosol particles are never spheres and the scattering functions present
large differences compared with equivalent spheres. However, Koepke and Hess (1988) showed that for aerosol types with a relatively low amount of large particles the effect due to uncertainty about particle shape can be ignored, compared to uncertain particle size and refractive index. Given the size distribution and complex refractive index the aerosols radiative properties can thus be determined from Mie calculations (see section 2.3.2) for continental and urban aerosol conditions. The phase functions present generally a large forward peak.

The aerosol optical properties in radiative transfer models are in most cases very roughly modelled, since the real size distribution and profile are unknown. The global CIE erythemal radiation decreases slowly for a large increase in the aerosol optical depth (Fig. 3.12, left and middle). However, the redistribution between direct and diffuse radiation is strongly influenced (Fig. 3.12, right). The large scattering in the forward direction leads to a redistribution of the direct radiation in diffuse downward radiation. In the example, the low turbidity (low aerosol amount) is defined by a total aerosol optical depth of 0.1 at 340 nm, and the high turbidity (high aerosol amount) by a total aerosol optical depth of 0.8 at 340 nm (inversely proportional to the first power of wavelength).
3.5 Effect of the surface albedo

Reflections of the radiation by the ground surface may increase the UV radiation measured at a station. The direct effect (local) of a reflection is the illumination of a downward-facing target, and the indirect effect (regional) is the illumination of the atmosphere from below, which can then scatter radiation back to the surface (Fig. 3.13). Usually, the surrounding region of about 10 km radius is considered to influence the radiation measured at a given place (Madronich, 1993). Note that the orientation of the target is important. Indeed, the radiation received perpendicular to the ground and facing the sun is much larger than back to the sun (also valid for a standing person).

3.5.1 Low and high surface albedo

The ground albedo depends on the surface properties (bare soil, vegetative, grass, rock, snow, water surface, snow, etc.). It usually takes different values for different wavelengths and zenith angles. For snow-free surfaces, the albedo in the UV is smaller than in the visible; the contrary is true for snow-covered surfaces. The albedo of a vegetative ground varies with time (during the year, for varying moisture conditions, etc.), but the surface with the most varying surface albedo is snow.

Measurements of the albedo in the UV range are presented in Blumthaler and Ambach (1988), Ambach and Eisner (1986), McKenzie et al. (1996),

![Figure 3.11: Typical number size distributions $-dN(r)/d\log r$ [cm$^{-3}$] of aerosol particles as a function of the radius [\mu m] of urban (solid curve), rural (dotted curve), maritime (dashed curve), and remote continental (dotted-dashed curve) models (Jaenicke, 1988)]
and Feister and Grewe (1995) (summary in Table 3.1). McKenzie et al. (1996) measured a spectral albedo over long grass between 1 and 2% in the UV range increasing slightly with the wavelength, and a small zenith angle dependency, especially in the UV-A. Some other results are also listed in Madronich (1993).

Snow on the ground is a multiple-scattering medium. The ice grains have generally optical sizes in the range 50 μm for new snow to 1 μm for old melting snow. Like cloud droplets, they are non selective scatterers of light. However, the scattering by the grains is more sharply peaked in the forward direction (g is close to 1) than scattering by cloud droplets. The ultraviolet
Table 3.1: Albedo of various surfaces [%] for the erythemal UV radiation and the total shortwave solar radiation. See references in text.

<table>
<thead>
<tr>
<th>Surface</th>
<th>Erythemal [%]</th>
<th>Total [%]</th>
</tr>
</thead>
<tbody>
<tr>
<td>Snow free</td>
<td>1-5</td>
<td>12-22</td>
</tr>
<tr>
<td>New dry snow</td>
<td>95</td>
<td>87</td>
</tr>
<tr>
<td>New wet snow</td>
<td>80</td>
<td>75</td>
</tr>
<tr>
<td>Old dry snow</td>
<td>82</td>
<td>79</td>
</tr>
<tr>
<td>Old wet snow</td>
<td>74</td>
<td>72</td>
</tr>
<tr>
<td>Old snow (dirty)</td>
<td>60</td>
<td>68</td>
</tr>
<tr>
<td>Liquid water</td>
<td>5</td>
<td>9</td>
</tr>
</tbody>
</table>

Light is less absorbed than visible and infrared. The snow albedo is larger for ultraviolet radiation than for visible and infrared because of nonselective multiple scattering in a selective absorbing medium. However, the optical properties of a snow covered surface depend very strongly on the moisture, the age and the cleanliness of the snow. One can find a good review of the optical properties of snow and of the radiative transfer models in snow in Bohren (1987) and Warren (1982).

3.5.2 Enhancement of the diffuse radiation

Although no natural surface is really lambertian (purely diffuse reflection), most of the radiative transfer models are based on the lambertian hypothesis (the DISORT algorithm can treat bidirectional reflectance). Some models use ever a mean albedo value for the whole considered spectrum, others make a wavelength dependent adaptation. The albedo is generally expressed as a fraction of 1 or as a percentage. The incident energy may involve direct and/or diffuse radiations.

The diffuse upward and downward erythemal UV radiation are significantly enhanced by the presence of a reflecting surface. The direct radiation is not influenced. In the example presented on Fig. 3.14, the enhancement of the diffuse downward radiation is sensitive up to about 10 km. This confirms the importance of the regional effect in the enhancement of the diffuse downward radiation.

3.6 Attenuation and enhancement by clouds

The clouds are one of the strongest parameters affecting the UV radiation reaching the ground. Clouds can both decrease and increase the UV radiation depending on their properties and positions. The microphysical (particle shape, size distribution, chemical composition) and macrophysical proper-
ties (shape of the cloud), as well as the interaction between clouds or between clouds and surface, determine the optical properties of a cloudy sky (Fig. 3.15). The modelling of the interaction between radiation and clouds is very complex. All the more that the microphysical properties of the clouds are usually unavailable. Many uncertainties remain on understanding the effect of clouds on the radiation. The discrepancy between measurements and radiation transfer calculations, especially absorption, are not understood (see e.g. Cess, 1995, Li and Moreau, 1996, and Kondratyev et al., 1998).

### 3.6.1 Cloud microphysics

The droplet in clouds form by heterogeneous nucleation on atmospheric aerosol particles (cloud condensation nuclei, CCN). As a consequence of their size and water solubility, only about 1% of continental air aerosol and about 10 - 20% of maritime air aerosol serve as CCN (Wallace and Hobbs, 1977). Typical cloud-drop size distribution can be described by a gamma distribution (Fig. 3.16). Further information about the formation and growing of the droplet and ice particles, as well as on the typical numbers of drops, water contents and concentrations of ice particles are given e.g. in Pruppacher and Klett (1997) and Wallace and Hobbs (1977). A summary of measurements of microphysical properties may also be found in Barthazy et al. (1997). General information about the cloud formation, classification and dynamics
3.6 ATTENUATION AND ENHANCEMENT BY CLOUDS

Figure 3.15: Important parameters in the cloud - radiation interaction: (A) chemical composition and particle shape; (B) shape of the cloud; (C) interaction between clouds; (D) interaction with the earth surface; (E) geometry sun - cloud(s) - observer.

may be found in Wallace and Hobbs (1977) or Houze (1993).

3.6.2 Attenuation and cloud-albedo interaction

The chemical composition define the refractive index of the particles. Assuming cloud droplets to be spherical water balls (a good assumption except for very large falling raindrops), the optical properties can be obtained from Mie theory (Savijärvi et al., 1997). Other phase functions are necessary to deal with cirrus clouds (crystals). Clouds droplets have scattering cross section that are essentially independent on the wavelength over the UV region ($x = 2\pi r/\lambda >> 1$). Absorption by pure water is negligible and impurities are usually too small to cause significant absorption, excepted in highly polluted regions. Clouds are thus expected to introduce only a weak wavelength dependence in the UV range. Indirect effects can, however, produce a wavelength dependence. Scattering within clouds, for instance, increases the photon path length through absorbing gases in the interstitial cloud space and below the cloud. Scattering of the radiation upwards from the cloud and downward again may also create a wavelength dependence in the transmission (Seckmeyer et al., 1996; Kylling et al., 1996).

In most radiation transfer models (except in Monte-Carlo calculations), a cloud is considered as a plane-parallel layer. The cloud optical depth is related to the liquid-water content (see e.g. Fouquart et al, 1989). Some authors give results about model calculations (see e.g. Madronich, 1987), but few about observations of the UV transmission in clouds (summary in Madronich, 1993). Transmission ranges from about 10% for large altocumulus clouds to more than 90% for cirrus clouds. The overall attenuation is in simple model a function of the fraction of the sky covered by cloud.
However, the results of such parametrization should be cautioned due to the high variability of cloud distributions and morphologies. Remarks about the effect of cloud shapes may be found in Welch and Wielicki (1984).

Almost no direct radiation remains below a cloud layer (Fig. 3.17), and the diffuse downward is substantially increased below the cloud relative to above the cloud. On the other hand, the diffuse upward above the cloud is much larger than without a cloud. The interaction between a high albedo surface and a cloud layer leads to an increase of the diffuse upward, especially under the cloud, and an increase of the diffuse downward under the cloud layer.

3.6.3 Enhancement due to reflection on clouds

The measured global radiation is sometimes exceeding the clear-sky value when the sky is partly cloudy and there is reflection off cloud sides into the field of view of the instrument. Enhancements of up to 30% were observed (see Mims and Frederick, 1994). The enhancement effects are highly variable and may or may not significantly increase surface irradiance. It is usually short in time, lasting up to an hour depending on cloud height, orientation of the sun, and prevailing wind speed. The increase is dependent on the amount of diffuse radiation scattered into the solid angle subtended by the projection of the cloud-side area normal to the sensor. The magnitude of cloud reflection events increases with increasing cloud height (the larger the area of the illuminated side, the larger the normal projection relative to the
Figure 3.17: Profile of the erythemal CIE radiation [W m$^{-2}$] with a cloud between 3 and 4 km ($g = 0.85, \omega_0 = 0.9999, \tau = 0.38$). Left: surface albedo is 0.03. Right: surface albedo is 0.5. Direct (solid thin curve), diffuse downward (dotted curve), diffuse upward (dotted-dashed curve), and global (solid thick curve) radiation. TUV calculations (sea level, zenith angle is 25°, ozone content is 300 DU, Elterman (1968) aerosol profile).

instrument). It will also increase with increasing zenith angle (fraction due to cloud geometry). However, at some critical zenith angle, cloud elements will begin to shadow one another, cutting off the surface from any direct component of the radiation and eliminating the possibility of enhancement (at low elevations angle, the cloud cover is always 1). Other factors affecting the enhancement magnitude include cloud optical depth, atmospheric aerosol concentration, and atmospheric column ozone (Harshvardhan, 1997).
Chapter 4

Data and radiative transfer model calculations

This chapter describes measurements of the solar radiation, ozone and meteorological parameters used in this work. The radiative transfer model used for numerical calculations is then presented with assumptions and remarks about the calculation procedure. Information about the influences of the orography on diffuse radiation at Davos can be found in the last section of this chapter.

4.1 Sites of measurements

Measurements at Davos, Payerne and Arosa represent the main data set for the analysis described in this work.

![Figure 4.1: Horizon at the station Davos (left) and Payerne (right). Apparent path of the sun in the sky June 21 (dashed curve) and December 21 (solid curve). Azimuth 0° pointing southward.](image)

The measuring site at Davos is situated at the Physikalisch-Meteorologisches Observatorium and World Radiation Center (PMOD/WRC) (46°48’ N, 9°49’ E;
1610 m a.s.l.). The station is located in a valley of the Grison mountains of eastern Switzerland. About 20% of the half hemisphere is hidden by the mountains (Fig. 4.1).

The measuring site at Payerne is located at the aerological station of the Swiss Meteorological Institute (SMI). The station is in the western plain of Switzerland (46°49' N, 6°57' E; 490 m a.s.l.), 220 km from Davos, with an almost free horizon (Fig. 4.1).

The station of Arosa is located at the Lichtklimatisches Observatorium (LKO) (46.78° N, 9.68° E; 1820 m a.s.l.). Arosa is in the eastern part of the Swiss Alps, about 12 km from Davos and is surrounded by mountains.

4.2 Solar radiation measurements

4.2.1 Erythemal UV radiation

The erythemal UV measurement network has been set up in the framework of the Swiss Atmospheric Radiation Monitoring program (CHARM). The instruments were installed by the Swiss Meteorological Institute (SMI; person in charge: A. Heimo) in collaboration with the PMOD/WRC (person in charge: R. Philipona) (Philipona et al., 1996).

Direct, diffuse and global components of the erythemal UV radiation are measured at Davos and Payerne since May 1995 by using UV-Biometers (Solar Light Co, model 501, Fig. 4.2) (Solar Light Co, 1991). The instrument for the direct radiation is in a special tube. This tube is mounted on a sun-tracking system and provides a viewing angle of 8 degrees. Diffuse radiation is measured with a fixed instrument (horizontal receiver) with a sun-tracking shading disk covering the sun by the same solid angle of 8 degrees. Global radiation is measured with a horizontally mounted instrument (without mask). Two-minute average values are recorded.

The global erythemal UV radiation is also measured at Arosa by using an "original" Robertson-Berger, a UV-Biometer (Solar Light Co., model 501), a UVB-1 ultraviolet pyranometer (Yankee Environmental systems Inc.; transferred to Tschuggen in April 1995), and a Macam PD104B-UVB. Furthermore, the direct and global erythemal radiation is measured at Jungfraujoch (46°33' N, 7°59' E; 3580 m a.s.l.) using UV-Biometers.

All these instruments are sensitive to a broadband UV range; their spectral responsivities are similar to the erythemal action spectrum defined by CIE (CIE, 1987). However, instruments have a slightly different spectral responsivity (see examples in Leszczynski et al., 1995), with differences obviously larger between different instrument types (for instance between the Robertson-Berger and the UV-Biometer) than within a given type.

The principle of the operation of the various instruments is similar. First,
the solar radiation incident on the quartz dome of the detector is filtered by a black prefilter that blocks the visible and near infrared radiation. Then, the remaining UV radiation is converted to visible radiation by a fluorescing phosphor layer deposited on a green post filter. The fluorescence radiation through the green filter that absorbs the residual UV radiation is detected by a silicon photo diode (Leszczynski et al., 1995). The UV-Biometer, as well as the UVB-1 ultraviolet pyranometer, are temperature stabilized. The instruments are generally calibrated relative to a spectrally integrated value weighted with the CIE curve. UV-Biometers yield MED/hour (1 mV/250 = 1 MED/hour), where 1 Med is the Minimal Erythema Dose (CIE, 1987) for the erythemal radiation with the relation 1 MED/hour = 5.83 $10^{-2}$ W/m² (Solar Light Co, 1991).

The UV-Biometers in the Swiss network are calibrated in Davos by comparison with the Swiss reference UV-Biometer (#1492) (person in charge: R. Philipona). The responsivity of this reference instrument (Fig. 4.3) was measured during the WMO/STUK intercomparison which took place in Finland in 1995 (results in Leszczynski et al., 1995). The calibration factor assures good accuracy for sun elevations larger than 30 degrees. The measurements for lower sun elevations need an additional correction. Accuracy and stability of the instruments are also checked regularly by comparison of the sum of the direct and diffuse radiation with the global radiation. Except some problems with the infiltration of humidity the instruments are quite stable in time (desiccant has to be changed regularly). The error in measurement is smaller than 5% in comparison with the Swiss reference. The absolute error is estimated to be smaller than 10% compared to the other European standard instruments (R. Philipona, personal communication).
4.2.2 Shortwave radiation and sunshine duration

The total shortwave radiation is measured at the automatic stations of the Swiss Meteorological Institute (ANETZ) using Kipp & Zonen CM6 pyranometers. Ten-minute average values are recorded (unit: W/m²) in the ENAD data base. The pyranometer measures the global solar irradiance. It has a glass dome to protect the blackened thermopile detector; this dome is transparent from approximately 300 nm to 2800 nm (Kipp & Zonen).

Ten-minute sunshine duration is recorded at the ANETZ stations using Haenni Solar IIIB instruments (unit: minute) with a threshold of 200 W/m².

4.2.3 Filter radiometer and aerosol optical depth

The spectral filter radiometer (sunphotometer) consists of interference filters and silicon detectors to measure direct solar radiation in a narrow spectral band (typically 5 nm). The instrument is calibrated in terms of its extraterrestrial signal at 1 AU distance from the sun. Measurements at 368, 412, 500, and 778 nm at Davos are used in this work. The calibration of the instruments is performed routinely at the PMOD/WRC by Langley plots (person in charge: Ch. Wehrli).

The aerosol optical depth $\tau_a$ at wavelength $\lambda$ can be derived from spectral filter radiometer data by using Beer-Bouguer-Lambert law (see chapter 2).

$$
\tau_a(\lambda) = -\frac{1}{m_a} \left( \log \left( \frac{V_{\lambda,0} d_e}{V_\lambda} \right) + m_p \tau_r(\lambda) + \mu \tau_o(\lambda) \right)
$$  (4.1)
where $V_\lambda$ is the measured irradiance and $d_c$ is the Sun-Earth distance correction (see Appendix A). The relative optical masses for air $m_p$, aerosol particle $m_a = m_r$, and ozone $\mu$ are described in Appendix B, as well as formula for the Rayleigh optical depth $\tau_r(\lambda)$ and the ozone optical depth $\tau_o(\lambda)$. The value of $V_{\lambda 0}$ is determined by the calibration procedure (Langley plot) and represents the irradiance that would be measured in extraterrestrial conditions. The influence of other absorbing gases ($NO_2$ and $SO_2$) can be neglected at Davos. Aerosol optical depths at 368 nm, 412 nm, 500 nm and 778 nm have been calculated at Davos (data received from Ch. Wehrli, PMOD/WRC).

### 4.3 Atmospheric ozone measurements

The total ozone amount has been measured by sunphotometry since 1926 at the Lichtklimatisches Observatorium (LKO) in Arosa. Since 1988, the Swiss Meteorological Institute is responsible for the operational performance of the ozone measurements. The series was homogenized and compared with satellite and other ground based measurements (Staehelin et al., 1998a; Staehelin and Renaud, 1996).

At the present time, total ozone is measured with two Dobson (D101, D62) and two Brewer (B40, B72) spectrophotometers (unit: DU). The results of the two Dobson instruments are given for the wavelength pairs AC, AD, C, and CD (see Staehelin et al. 1995, for the description of the measurement method). Direct solar measurements are available for 200-300 days per year, with up to about 10 single measurements per day. If not specified otherwise, the D101(AD) measurements are used in this work.

Ozone balloons have been launched in Payerne since 1969, at the frequency of three ascents per week. Brewer-Mast sondes are used to determine the ozone concentration profiles (unit: nbar). See Staehelin and Schmid (1991), and Stübi et al. (1996) for more information about the soundings and homogenization of the series.

### 4.4 Meteorological data

The standard surface meteorological parameters (temperature, humidity, dew point, wind direction and velocity, etc.) are measured in Switzerland at ANETZ stations. Ten-minute average values are recorded. Data from the stations Davos, Weissfluhjoch (2690 m a.s.l., ca 6 km from Davos) and Payerne are used in this report (ANETZ stations close to the radiation instruments).

The prevailing weather is observed regularly for forecasting purpose. International code tables of the WMO are used for the surface synoptic mes-
sages. In Payerne, SYNOP observations are made every 3 hours with the description of the cloud type and coverage of the sky for the three altitude groups. In Davos, observations are conducted at 6, 12 and 18 UTC with the description of the cloud coverage but not of the detailed cloud types. The horizontal visibility is also recorded at Payerne and Davos.

Meteorological radiosondes are launched every day at 00 and 12 UTC from Payerne. They measure the profile of the temperature, relative humidity, dew point, wind direction and velocity in the atmosphere.

The European Center for Medium Range Weather Forecast (ECMWF) performs analysis and forecast of meteorological fields over Europe. These data are available at the Institute for Atmospheric Science (LAPETH).

4.5 Radiative transfer model calculations

The Tropospheric Ultraviolet-Visible (TUV) model developed by S. Madronich is used for radiative transfer calculations (program available by anonymous ftp to sasha.acd.ucar.edu). The model uses a multi-layer delta-Eddington radiative transfer scheme (Joseph et al., 1976) with a fast tri-diagonal matrix solution (Toon et al., 1989), or a discrete ordinate method (Stamnes et al., 1988) with 8 streams (may be changed). The calculation of UV radiation reaching the ground is carried out over 280-400 nm at 1 nm intervals. The atmosphere is divided into 50 layers of 1 km height each (may be changed). TUV is time efficient and was used for many applications (e.g. Madronich, 1987).

In general, we assume that the results are accurate to about 10% for the two-stream method and better for DISORT (better modelling of the scattering processes). Larger errors may occur at large zenith angles or for large quantities being considered. Accuracy and uncertainties in modeled UV irradiance are discussed e.g. in Weihs and Webb (1997a) and Schwander et al. (1997). Comparisons between two-stream and DISORT can be found in Piers et al. (1993) and Kylling et al. (1995). The two-stream algorithm of TUV has also been shown to agree well with more sophisticated multiple scattering models (larger number of streams) with a wide range of mid-latitude atmospheres (Koopke et al., 1998).

4.5.1 Input parameters and assumptions

The extraterrestrial solar irradiance originates from SUSIM SL2 measurements (van Hoosier et al., 1987) for wavelengths smaller than 350 nm, and Neckel and Labs (1984) for wavelengths larger than 350 nm. The ozone absorption cross section for wavelengths between 240.5 and 350 nm is from Molina and Molina (1986) with temperature adaptation. The ozone absorp-
tion cross section for wavelengths larger than 350 nm is from WMO (1985b). The O₂ cross section is from Brasseur and Solomon (1986). The absorption cross section of NO₂ is from Davidson et al. (1988), and the absorption cross section of SO₂ is from McGee and Burris (1987).

The optical properties of the atmosphere at each height (profile) are needed to compute the radiative transfer. Since many optical properties are unknown for a given time and place, standard profiles have been defined. Vertical profiles of temperature and air density are from the US Standard Atmosphere at mid-latitude (1976). If not specified, the basic vertical ozone profile is from the US Standard Atmosphere at mid-latitude (1976). The ozone concentrations are rescaled by the total ozone amount in equal proportion at each height. The variation in the ozone profile pattern will be studied in chapter 5.

We consider an atmosphere without NO₂ and SO₂ (valid assumption for the considered cases). The surface albedo is assumed equal to 0.03 for surfaces not covered by snow. It remains independent on the wavelength with the assumption of Lambertian reflection.

The aerosol optical properties are defined by the profiles of the optical depth τₐ, asymmetry parameter g, and single scattering albedo ω₀. The aerosol optical depth profile of Elterman (1968) is used as a basis for aerosol optical properties (data for 340 nm). The aerosol optical depths are then rescaled by the total aerosol optical depth at 340 nm (scaling implemented). If not specified, the aerosol optical depth scales inversely with first power of wavelength (λ⁻¹). Additional wavelength dependencies as well as variations in the profile will be discussed in chapter 6.

In this work, the input parameters for clear-sky conditions are total ozone amount, total aerosol optical depth, asymmetry parameter and single scattering albedo of aerosol particles, surface albedo and altitude. In case of a cloudy sky, the cloud optical depth, asymmetry parameter, single scattering albedo, top and base of cloud layers are added. All other parameters remain constant. Note that the various input profiles, such as aerosol and standard atmosphere, are rather simple. For instance, the mean ozone profile changed during the last decades. However, these differences should not have a large influence on the results presented in this work (see sensibility calculations in next chapters).

The program TUV has been adapted to obtain different output files (spectral irradiances, erythemal integrated value as a function of the zenith angle, profile of the erythemal integrated value). The erythemal UV radiation is calculated for the CIE action spectrum (UV_{CIE}) and for the responsivity of the UV-Biometer Swiss reference (UV_{CH}, Fig. 4.3). Direct, diffuse downward and diffuse upward radiation is calculated in any case.

Uncertainties remain in calculations for large zenith angles. Therefore, comparisons with UV-Biometer measurements (also less reliable for large
zenith angles) are only carried out for zenith angles smaller than 70°.

4.5.2 Calculation of the erythemal UV radiation

The UV-Biometer has a spectral responsivity which differs from the CIE action spectrum (Fig. 4.3). As a result, the ratio between CIE and UVBio(CH) integrated values $\frac{U_{V_{\text{CIE}}}}{U_{V_{\text{CH}}}}$ depends on the zenith angle and ozone content (Fig. 4.4).

The UV-Biometers are calibrated by comparison with the CIE integrated irradiance, for given total ozone amount and zenith angles. As a result of shifts in the spectrum, UV-Biometer measurements for other atmospheric conditions and altitude are not directly comparable with the CIE integration\(^1\) (see also Mayer and Seckmeyer, 1996). The remark is also valid for the direct and diffuse radiation.

Since the ratio between CIE and UVBio(CH) integrated irradiance at Davos equals about 0.5 for zenith angles smaller than 60°, comparisons between UV-Biometer measurements and model calculations are presented for $U_{V_{\text{CIE}}}$ and 0.5 $U_{V_{\text{CH}}}$ in this work\(^2\).

![Figure 4.4: Ratio between the CIE and UVBio(CH) weighting radiations, as a function of the zenith angle for 250 DU (solid line), 300 DU (dotted line) and 450 DU (dashed line). TUV calculations (altitude of Davos and Elterman (1968) aerosol profile).](image)

\(^1\)The ratio between CIE and UVBio(CH) integrated irradiance equals 10.8/11.9 = 0.91 for extraterrestrial solar spectrum and about 0.5 at Davos for zenith angles smaller than 60°.

\(^2\)The 0.5 factor is due to differences in relative sensitivities and in the fact that the CIE and UV-Biometer sensitivities are scaled to 1 at 300 nm. Note that $\int_{290 \text{ nm}}^{400 \text{ nm}} p_\lambda \, d\lambda$ equals about 13 when $p_\lambda$ is the CIE action spectrum, and 15.5 when $p_\lambda$ is the responsivity of the Swiss reference UV-Biometer, respectively.
4.6 Diffuse radiation and orography

Davos is in a valley surrounded by mountains. Measurements are therefore not directly comparable with radiative transfer calculations, which assume a flat horizon. In this section, we estimate the effect of the orography on the radiation measured at Davos.

4.6.1 Snow free surface and clear-sky

Global radiation on days with a snow free surface at Davos $glo_{dav} (no\; snow)$ and in a flat land $glo_{flat} (no\; snow)$ can be decomposed as followed

$$glo_{dav} (no\; snow) = dir + dif(sky\; 1) + dif(sky\; 2)$$
$$glo_{flat} (no\; snow) = dir + dif(sky\; 1) + dif(sky\; 2) + dif(sky\; 3)$$

where $dir$, $dif(sky\; 1)$, $dif(sky\; 2)$ and $dif(sky\; 3)$ are as illustrated on Fig. 4.5. The part of radiation in $dif(sky\; 3)$ depends on the wavelength, on the zenith angle, on the atmospheric optical properties and on the horizon (defined by azimuth and elevation). It may be estimated by using a clear-sky radiance distribution.

Grant et al. (1997a) showed that the best empirical radiance distribution fitting their measurements in the UV-A, UV-B and total shortwave ranges was the normalized radiance $N \; [W \; m^{-2} \; sr^{-1}]$ given for the direction of zenith angle $\theta$ and scattering angle $\psi$ (angle between the sun and the considered point).

$$N(\psi, \theta) = A + \frac{B \theta^2}{\pi/2} + C \; e^{-m\psi} + D \; \cos^2 \psi$$

The scattering angle $\psi$ is defined by $\cos \psi = \cos \theta \cos \theta^* + \sin \theta \sin \theta^* \cos \phi$, where $\theta^*$ is the solar zenith angle and $\phi$ is the azimuth between the sun and
the considered point. The parameters \( A, B, C, D \) and \( m \) have to be estimated (values from Grant et al., 1997a for the UV-A and UV-B ranges in Table 4.1). The term \( A \) in Eq. 4.4 represents the isotropic sky or multiple scattering background, the second term describes horizon-brightening component, and the combined third and fourth terms represents the circumsolar scattering component. Examples of sky distribution are shown in Grant et al. (1997a) for UV-A, UV-B and total shortwave, and in Ireland and Sacher (1996) for erythemal UV radiation.

Table 4.1: Coefficients of the least square fit to clear sky radiance distribution, from Grant et al. (1997a). Note that the region within 15° of the solar position (solar disk) is excluded.

<table>
<thead>
<tr>
<th></th>
<th>( A )</th>
<th>( B )</th>
<th>( C )</th>
<th>( D )</th>
<th>( m )</th>
</tr>
</thead>
<tbody>
<tr>
<td>UV-A</td>
<td>0.20</td>
<td>0.12</td>
<td>2.7</td>
<td>0.14</td>
<td>7.6</td>
</tr>
<tr>
<td>UV-B</td>
<td>0.19</td>
<td>0.03</td>
<td>1.4</td>
<td>0.14</td>
<td>7.8</td>
</tr>
</tbody>
</table>

The horizon at Davos has an elevation between 2 and 24° with an average of 12° (Fig. 4.1). In the estimation of the diffuse part hidden by the mountains, we assume that the elevation at Davos is 12° for all azimuths. Based on the parameters in Tab. 4.1, the integrated irradiance \( \int \int N(\psi, \theta) \cos \theta \sin \theta \, d\theta \, d\phi \) was estimated in the UV-B and UV-A ranges with and without mountains. The component \( \text{dif} \text{(sky3)} \) for the erythemal UV radiation corresponds then to about 2 - 5% of the whole erythemal diffuse radiation, varying with the wavelength range and zenith angle (Table 4.2).

Table 4.2: Estimation of the attenuation [%] in diffuse UV-A and UV-B irradiance due to the mountains at Davos. Clear-sky radiance distribution from Grant et al. (1997a). \( \theta^* \) is the solar zenith angle.

<table>
<thead>
<tr>
<th></th>
<th>( \theta^* = 25° )</th>
<th>( \theta^* = 70° )</th>
</tr>
</thead>
<tbody>
<tr>
<td>UV-A</td>
<td>-2.96%</td>
<td>-4.86%</td>
</tr>
<tr>
<td>UV-B</td>
<td>-2.0%</td>
<td>-3.68%</td>
</tr>
</tbody>
</table>

Note that these estimations are valid for low surface albedo and clear sky. The parameters in Eq. 4.4 were estimated by Grant et al. (1997a) in a clean continental atmosphere and low altitude. The radiance distribution in Davos is then probably less intense. The calculations above show anyway that the orography at Davos has in most cases only a small impact on the diffuse radiation. This impact should not be very important in comparison with other uncertainties; with exception maybe for large zenith angles when part of the 15° circumsolar radiation can be hidden.
4.6.2 Snow covered surface and clear-sky

Similar to the decomposition for surface not covered by snow, the radiation measured at Davos \( g_{dav}(snow) \) and in a flat land \( g_{flat}(snow) \) with a snow covered surface can be decomposed as followed:

\[
g_{dav}(snow) = g_{Dav}(no\ snow) + dif(surf\ 1) + dif(surf\ 2) + A_1 dif(sky\ 3) + A_2 dif(surf\ 3) \tag{4.5}
\]

\[
g_{flat}(snow) = g_{flat}(no\ snow) + dif(surf\ 1) + dif(surf\ 3) + A_3 dif(surf\ 2) \tag{4.6}
\]

where \( dif(surf\ 1), dif(surf\ 2) \) and \( dif(surf\ 3) \) are illustrated on Fig. 4.6, and \( g_{dav}(no\ snow) \) and \( g_{flat}(no\ snow) \) given in equations 4.2 and 4.3. Radiation reflected by the mountainsides \( dif(surf\ 2) \) is the only part really influencing the global radiation by direct reflection of diffuse and direct radiation into the field of view. The part of radiation reflected from behind the mountains is given by the expression \( A_1 dif(sky\ 3) + A_2 dif(surf\ 3) \) and is not negligible as the diffuse radiation is increased by the albedo from a circle of at least 10 - 20 km radius (see section 3.5).

![Figure 4.6: Illustration of the main interactions between the radiation and the mountains for a snow covered surface. Components in Fig. 4.5 are not repeated.](image)

The difference of radiation due to the orography equals then

\[
g_{flat}(snow) - g_{dav}(snow) = (1 - A_1) dif(sky\ 3) + (1 - A_2) dif(surf\ 3) - (1 - A_3) dif(surf\ 2) \tag{4.7}
\]

The component \( (1 - A_1) dif(sky\ 3) \) is smaller than 2 - 5% of diffuse erythermal radiation (see above), and the component \( (1 - A_2) dif(surf\ 3) \) is probably much smaller than 2 - 5% as \( dif(surf\ 3) \) is smaller than \( dif(sky\ 3) \). Component \( (1 - A_3) dif(surf\ 2) \) may be very large depending on the geometry and on the surface and atmosphere optical properties.
Thus the decrease of diffuse radiation due to the orography at Davos is expected to be smaller with a snow covered surface than with a snow free surface (values larger at Davos than in a flat land are also theoretically possible) and depends on the albedo, orography, wavelength, zenith angle and atmospheric properties. Note that if the reflection is lambertian, the fact that the mountains have slopes does not influence the albedo effect. The orography could however enhance a natural non-lambertian behaviour of surface reflection (see discussion in chapter 7).

4.6.3 Cloudy sky

Grant and Heisler (1997) and Grant et al. (1997b) gave results about the ultraviolet radiance distributions in case of an obscured, respectively translucent, overcast sky. They concluded that radiance in the UV-B, UV-A and visible ranges with obscured overcast skies can be modeled as a function of the zenith angle of the radiance. The radiance distribution for a translucent overcast sky was modeled with a similar formula as in clear-sky conditions (see Eq. 4.4). The orography influence on overcast days at Davos is expected to be smaller than in the clear sky case, as the contribution from the mountains solid angle is about 2-3% of the hemisphere.
Chapter 5

Ozone and zenith angle

In this chapter, we present results about the influences of ozone and zenith angle on the erythemal UV radiation. Only clear-sky conditions are considered. Statistical models based on clear-sky UV-Biometer measurements are described. These models allow comparing total ozone and zenith influences, as well as estimating effect of variations in total ozone amount at given zenith angle.

A normalization procedure to constant total ozone amount and constant Sun-Earth distance is presented in the last section of this chapter. This procedure is applied in the following chapters. It filters out the effect of varying total ozone amounts and Sun-Earth distance for the isolation of other influences such as aerosol, surface albedo and clouds.

5.1 Clear-sky models at Davos and Payerne

Daily maximum measurements vary strongly during the year mostly because of changing solar zenith angle (Fig. 5.1). In clear-sky conditions, the day-to-day variability at a given station is mainly due to variations in the atmospheric ozone amount (Fig. 5.2).

The influence of total ozone and solar zenith angle on clear-sky global erythemal UV radiation is determined by using a statistical model

\[ UV_{Bio}(X, \theta, d_c) = d_c \cos \theta a_0 e^{a_1 X + a_2 m_r} e^{a_3 m_r^2 + a_4 (X \mu)^2 + a_5 X \mu m_r} \]  

(5.1)

where \( UV_{Bio}(X, \theta, d_c) \) is the global UV-Biometer data [W/m²] for total ozone amount \( X \) [DU] and at solar zenith angle \( \theta \). The variable \( d_c \) is the Sun-Earth distance correction (see Appendix A), \( \mu \) is the ozone relative optical mass (layer at 23 km above sea-level, see Appendix B), and \( m_r \) is the relative optical air mass for molecules (see Appendix B). Eq. 5.1 is based on the Beer-Bouguer-Lambert law, which is valid for the direct monochromatic radiation. Adaptations have been then included to fit the global UV-Biometer data (e.g. quadratic terms).
We estimated the unknown parameters $a_0-a_5$ by using a multiple regression method (see Appendix C) on clear-sky data. Clear-sky data were first identify by using sunshine duration data at the nearby ANETZ station. A correction is then applied to extract remaining cloud influences in order to really have clear days. This correction is based on a minimal ratio direct/global or minimal pyranometer value dependent on the zenith angle. See the remarks about the relation between cloud coverage and sunshine duration in chapter 8.

### 5.1.1 Clear-sky models for Davos

The clear-sky model for Davos was fitted with global UV-Biometer measurements between May 95 and April 96. We assumed the air mass $m_T$ (Kasten,
1966; see Appendix B) independent on the surface pressure\(^1\). As the erythemal radiation is strongly modified by the surface albedo (see section 3.5), we redistributed the clear-sky data into snow and no-snow clear-sky data.

No-snow data are measurements on days without snow at Davos and Weissfluhjoch (2690 m a.s.l., about 6 km from Davos). The criterion includes data at Weissfluhjoch to avoid the influence of large albedo in the neighbourhood (regional effect). The snow data are measurements on days with snow at Davos and Weissfluhjoch. The no-snow model was estimated with the no-snow clear-sky data, and the snow model with the snow clear-sky data (Table 5.1). Only data with zenith angles smaller than 70° were considered in the regression analysis, as data are less reliable for larger zenith angles.

Table 5.1: No-snow and snow clear-sky models at Davos. Parameters \(a_0\) - \(a_5\) with standard error, number \(N\) of data used in the estimation, ranges of total ozone amount \(X\) and zenith angle \(\theta\).

<table>
<thead>
<tr>
<th></th>
<th>No-snow model</th>
<th>Snow model</th>
</tr>
</thead>
<tbody>
<tr>
<td>(a_0)</td>
<td>(\exp(0.4905 \pm 1.124 \times 10^{-2}))</td>
<td>(\exp(0.3148 \pm 1.365 \times 10^{-2}))</td>
</tr>
<tr>
<td>(a_1)</td>
<td>(-3.542 \times 10^{-3} \pm 9.544 \times 10^{-5})</td>
<td>(-3.627e-3 \pm 4.286 \times 10^{-5})</td>
</tr>
<tr>
<td>(a_2)</td>
<td>(-0.7617 \pm 2.127 \times 10^{-3})</td>
<td>(-0.4036 \pm 1.735 \times 10^{-2})</td>
</tr>
<tr>
<td>(a_3)</td>
<td>(0.2402 \pm 2.859 \times 10^{-2})</td>
<td>(1.412 \times 10^{-2} \pm 7.984 \times 10^{-3})</td>
</tr>
<tr>
<td>(a_4)</td>
<td>(2.775 \times 10^{-6} \pm 4.034 \times 10^{-7})</td>
<td>(7.347 \times 10^{-7} \pm 6.399 \times 10^{-8})</td>
</tr>
<tr>
<td>(a_5)</td>
<td>(-8.541 \times 10^{-4} \pm 2.111 \times 10^{-4})</td>
<td>(2.864 \times 10^{-4} \pm 3.965 \times 10^{-5})</td>
</tr>
<tr>
<td>(N)</td>
<td>7318</td>
<td>7203</td>
</tr>
<tr>
<td>(X)</td>
<td>240 - 330 DU</td>
<td>250 - 420 DU</td>
</tr>
<tr>
<td>(\theta)</td>
<td>23 - 70°</td>
<td>28 - 70°</td>
</tr>
</tbody>
</table>

Total ozone amount and zenith angle explained 98% of the variance of the UV-Biometer data. Extremes of overestimation (negative residuals) as well as some of the extreme underestimations (positive residuals) are due to cloud influence (see Table 5.2 and examples below).

A test on an independent data set (May - December 1996) showed similar results for the snow model. The relative error ranged between -20% and 7.5%, and 3.7% of the absolute residuals were larger than 10%. The results for the no-snow model, however, exhibited a systematic overestimation of about 5%. About 10% of the absolute residuals were larger than 10%. Since no calibration problem has been detected, this systematic overestimation might be due to systematic differences between atmospheric conditions during the analysis period and the test period.

The clear-sky models for Davos (no-snow and snow models) give satisfactory results in most cases. However, some of the variations observed in the

\(^1\)No improvement was found when introducing the surface pressure.
Table 5.2: Accuracy of the no-snow and snow clear-sky models at Davos (May 1995 - April 1996).

<table>
<thead>
<tr>
<th></th>
<th>No-snow model</th>
<th>Snow model</th>
</tr>
</thead>
<tbody>
<tr>
<td>Relative residual</td>
<td>[-41%, 20%] with standard deviation 4.9%</td>
<td>[-27%, 16%] with standard deviation 4.5%</td>
</tr>
<tr>
<td></td>
<td>no dependence on ozone and zenith angle</td>
<td>no dependence on ozone and zenith angle</td>
</tr>
<tr>
<td></td>
<td>Relative residuals</td>
<td>average 3.5%</td>
</tr>
<tr>
<td></td>
<td>4% of the residuals are larger than 10%</td>
<td>2% of the residuals are larger than 10%</td>
</tr>
</tbody>
</table>

UV-Biometer measurements are badly fitted by the models.

As an example, the underestimation on August 9, 1995 (Fig. 5.3) is clearly due to cloud influence, possibly from enhancement by reflection. The difference in the accuracy between October 19 and 20, 1995 is probably due to a change in the atmospheric turbidity. Actually, we observed a larger part of direct radiation on 20.

The underestimation on February 24, 1996 (Fig. 5.4) can be partially explained by the decrease in the ozone amount during the morning. A surface albedo larger than average and/or a turbidity smaller than average may also be a reason for the observed underestimation. The asymmetric underestimation on March 10, 1996 is explained by changes in the turbidity. The ratio direct/global is clearly larger on 10 than on 9 between 10 and 13:30 UTC, and shows the same asymmetry pattern.

5.1.2 Air mass and ozone profile

The relative optical masses $m_\tau$ and $\mu$ are functions of the zenith angle. They base on assumptions about the air molecules and ozone concentrations profiles, and therefore could add some uncertainties in the model.

The air mass $m_\tau$ from Kasten (1966) should be reliable since the air molecule properties may be considered constant. The pressure in Davos varies by about $\pm 2 - 3\%$ around the average. That means variation of maximum $\pm 2\%$ for the erythemal UV-Biometer radiation during the year. Note that Koepke et al. (1996) estimated differences smaller than 3% from calculations at sea level with pressure varying from 950 hPa to 1040 hPa (reference at 1013 hPa, 330 DU total ozone amount, 0.25 aerosol optical depth at 550 nm).
A change of 1% in ozone mass $\mu$ has the same consequence as a change of 1% in total ozone amount in the clear-sky models. The parameterization of $\mu$ has then to be done carefully. The total ozone amount measured at LKO at solar zenith angle $\theta$ is reduced to the vertical direction with the assumption of a thin layer of ozone at 23 km altitude. As we use the total ozone values from LKO, the same ozone mass $\mu$ (Appendix B) should be accurate in our clear-sky models.

The profile of the atmospheric ozone varies during the year. Variations in the erythemal radiation are thus also expected (see section 3.3). We performed radiative transfer calculations for a winter profile and a summer profile (December 31, 1996 and July 7, 1995, see Fig. 3.8) (profiles both scaled to same total ozone amount, Elterman (1968) aerosol). As expected, the direct radiation does not depend on the profile pattern. On the other hand, the diffuse radiation was 4% larger with the winter profile than with the summer profile for zenith angle 25°. This value decreased for zenith angles increasing up to 70° (<1%) and increase again for zenith angles larger than 70°. The difference of global radiation was smaller than 3% for zenith angles between 25 and 70°. These results confirm that tropospheric ozone is
Figure 5.4: UV-Biometer measurements (solid curve) and snow clear-sky model (dashed curve) at Davos. December 1, 1995 (293 DU), February 24, 1996 (328 DU), March 9, 1996 (375 DU) and March 10, 1996 (345 DU).

more efficient (absorbs more radiation) than stratospheric ozone. Contrary to results given in Brühl and Crutzen (1989) or Madronich (1993), however, the summer profile remains more efficient for all zenith angles, maybe due to the uncertainties in the large zenith angles.

Note also that the direct radiation measured with solar zenith angle 70° went through an ozone layer (23 km altitude) located at above 67 km away from the observer. This can be probably neglected in most cases; however, it could have a significant influence on days with large spatial gradients in the stratospheric ozone concentrations.

As a conclusion, we may assume that variations in erythemal UV-Biometer radiation due to variations in pressure and ozone profile remain small in comparison with uncertainties in measurement, turbidity and surface albedo.

5.1.3 Clear-sky model for Payerne

A clear-sky model was developed for the station at Payerne (490 m a.s.l.). Some adaptations were added to the method used for Davos. First, surface
5.1. CLEAR-SKY MODELS AT DAVOS AND PAYERNE

pressure $p$ was included in the calculation of the air mass (see Appendix B). And second, a correction based on the ozone soundings was added to the total ozone measurements at Arosa to get the total ozone amount at Payerne. The model explained 96.7% of the UV-Biometer variance with parameters $a_0 - a_4$ ($a_5$ was not statistically significant). The standard deviation of the residuals equalled 8%. The average of the absolute deviations was 6.4%. Details about the procedure and results can be found in Röösli (1997).

Note that a redistribution of total ozone amount into stratospheric and tropospheric ozone amounts would probably increase the accuracy of the model at Payerne, especially in summer months, as both parts have different influences on erythemal radiation.

5.1.4 Remarks on the clear-sky models

The model described by Eq. 5.1 is similar to the model developed by Burrows et al. (1994) for forecasting UV in Canada. Burrow’s model was calibrated with clear-sky measurements at Toronto during 1989 - 1991 and explained the data with a standard deviation of about 8%. The fit accuracy was then similar for Payerne and Toronto, and better at Davos (standard deviation smaller than 5%). The reason for this difference is probably the lower turbidity and thus the lower variation at higher altitudes. The residuals at Payerne are indeed correlated with the ratio direct/global of the UV-Biometer radiation; which is not the case for Davos. Thus the aerosol content seems to have a significant influence on the global radiation in Payerne.

Empirical models are usually found to be less sensitive to ozone variations than radiative transfer models do (see e.g. Koepke et al., 1998). The main reason is probably that the range of ozone measurements is not broad enough to get good estimates for small (few measurements) as well as for large ozone values. Koepke et al. (1996) calculated a relative difference in erythemal UV irradiance equal to about +35% for 260 DU and -30% for 450 DU against 330 DU (sea level, 0.25 aerosol optical depth at 550 nm). For the clear-sky model at Davos a change from 260 to 330 DU results in a UV change of +24% to +32% for the no-snow model, and from +26% to +36% for the snow model (increasing with the solar zenith angle). For a change from 440 to 330 DU, the snow model yields -28% to -31%. We have then very similar dependence on variations in total ozone amount.

The no-snow and snow models take into account the variations in the zenith angle, total ozone amount and two surface albedo (snow, no-snow). Variations in other parameters (e.g. aerosol) affect the fit accuracy. Note also, that the clear-sky models are valid for the total ozone and zenith angle ranges used in the estimation of the parameters (legal ranges). The results from an extrapolation to smaller or larger ozone amount, as well as smaller or larger zenith angles should be cautioned.
5.2 Ozone and zenith angle dependency

Variations in total ozone amount have an important impact on the erythemal UV radiation reaching the ground. However, the ozone effect remains relatively small in comparison with the zenith angle effect (Fig. 5.5). The erythemal UV radiation for typical zenith angles and total ozone amounts on the 15th of each month is also showed on Fig. 5.5. We note the largest values in summer and the smallest in winter (zenith angle influence). The values are also larger in fall than in spring because of the smaller total ozone amounts (see Fig. 3.6). Note, however, that the ozone amount has a strong day-to-day variation in spring.

\[ \text{RAF} = - \frac{\Delta UV}{UV} \times \frac{\Delta X}{X} \] (5.2)

where \( UV \) and \( \Delta UV \) are erythemal UV radiation and its change, and \( X \) and \( \Delta X \) are the total ozone amount and its change. The linear approach is only valid for very small ozone changes. Booth and Madronich (1994) proposed an expression to describe a somewhat wider range of ozone change

\[ \text{RAF} = \frac{\log\left(\frac{UV_2}{UV_1}\right)}{\log\left(\frac{X_2}{X_1}\right)} \] (5.3)
where \( UV_1 \) and \( UV_2 \) are erythemal UV radiation corresponding to ozone amounts \( X_1 \) and \( X_2 \). In the following, we consider the increase of erythemal UV radiation for a decrease of 1\% total ozone amount.

Typical RAFs for the erythemal UV radiation are found to be around 1\% with maximal values of 2\% (Feister, 1994; Booth and Madronich, 1994; McKenzie et al., 1991). They are larger for zenith angle of 60\° than for zenith angle of 0\°.

The increase in global erythemal UV-Biometer radiation at Davos for a decrease of 1\% in ozone ranges between 0.8\% and 1.5\% with average 1.05\% over the legal range of the no-snow model (Fig. 5.6). The standard deviations equals 2 - 3\% of the RAF value for the larger part of the ozone and zenith angle ranges, but reaches 11\% for large ozone amount and large zenith angle. RAFs for the snow model have a slightly higher average value (1.1\%) and a similar dependence on the ozone and zenith angle.

**Figure 5.6**: Increase in global UV-Biometer measurement for a 1\% decrease in total ozone. No-snow model.

We observe two complementary behaviours. The effect on radiation of a 1\% ozone decrease first increases exponentially with the zenith angle for small ozone amounts. This behaviour can be explained by the exponential behaviour of the Beer-Bouguer-Lambert law. Furthermore, the effect on radiation decreases with increasing ozone amount and for large zenith angles. This is mainly explained by large ozone absorption. Almost no UV-B radiation reaches the ground for large zenith angles and large ozone amounts. A variation in ozone amount has therefore no more influence on the erythemal radiation.

RAFs have to be used very carefully. A variation in the spectral sensitiv-
5.3 Trend in UV irradiance?

Over the last twenty years, there has been some concern that biologically active ultraviolet radiation at the earth's surface may increase due to reduction in the amount of ozone in the stratosphere.

5.3.1 Calculated and measured trends

Most of the estimations about possible trends in erythemal UV radiation come from radiative transfer calculations based on measured ozone trends (see examples in Tab. 5.3). Some of them take into account both tropospheric and stratospheric ozone trends (e.g. Madronich, 1993), and the tropospheric aerosol and cloud effects (e.g. Herman et al., 1996). According to Liu et al. (1991) the observed visibility reduction due to aerosol particles may account for reductions of 5 - 18% in UV-B since the industrial revolution (non-urban area). The determination of trends in stratospheric as well as tropospheric ozone remains however uncertain because of data reliability and variability.
5.3. TREND IN UV IRRADIANCE?

(see e.g. Staehelin et al., 1998a and 1998b).


<table>
<thead>
<tr>
<th>Basis</th>
<th>Period</th>
<th>Increase per decade</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>TOMS, 45° N</td>
<td>1979 - 1989</td>
<td>+ 4 ± 1.4%</td>
<td>(1)</td>
</tr>
<tr>
<td>Payerne soundings winter</td>
<td>1967 - 1990</td>
<td>+ 4.1%</td>
<td>(2)</td>
</tr>
<tr>
<td>Payerne soundings summer</td>
<td>1967 - 1990</td>
<td>+ 1.4%</td>
<td>(2)</td>
</tr>
<tr>
<td>TOMS, 45°± 5° N, April</td>
<td>1979 - 1992</td>
<td>+ 5.1%</td>
<td>(3)</td>
</tr>
<tr>
<td>TOMS, 45°± 5° N, summer</td>
<td>1979 - 1992</td>
<td>+ 3.6%</td>
<td>(3)</td>
</tr>
</tbody>
</table>

Few measurements of erythemal radiation are available for the past two decades. Furthermore, the interpretation of trends is complicated by the difficulty of maintaining calibrations of instruments over long periods of time, by the different sensitivity to ozone depletion (instrument responsivity), and by the variability and possible trends in other parameters such as clouds, surface albedo and aerosol content and type. As an example, the paper of Kerr and McElroy (1993) and the comments of Michaels et al. (1994) show the difficulty in the interpretation of estimated trends at 300 nm at Toronto. The choice of the considered period and of the data (single measurements or daily total, clear-sky or all-weather data, etc.) may lead to very different values.

No significant positive trend has been measured in regions with high densities of population. Scotto et al. (1988) reported from Roberston-Berger instruments an UV decrease of 5% to 11% per decade from 1974 to 1985 in the USA. However, Weatherhead et al. (1997) noted that problems with the US network rendered the existing measurements unreliable for long-term trend analysis. Different treatments of the data resulted in dramatically different trend results. The only significant positive trends were found in the high altitude station Jungfraujoch (3576 m a.s.l.) for clear-sky days. Trend estimations showed a 11 ± 4% increase per decade from 1981 to 1989 (Blumthaler and Ambach, 1990), and 7% ± 3% per decade from 1981 to 1991 (Blumthaler, 1993).

5.3.2 Theoretical erythemal UV increase at Davos

The largest monthly average total ozone depletion at Arosa is observed in March (Staehelin et al., 1998b). The total ozone trend was -4.6 ± 1.2% per decade from 1970 to 1997. The corresponding theoretical increase in erythemal UV-Biometer radiation would be +5.2 ± 1% per decade on March 15
at 12:00 (solar time) at Davos. On the other hand, the theoretical increase of UV-Biometer radiation on June 15 at 12:00 (solar time) would be +2.4 ± 0.4% per decade, as total ozone amount was decreasing -2.4 ± 0.5% per decade during the same period (Fig. 5.8).

![Graph showing monthly average total ozone amount (solid line) at Arosa during March (above) and June (below) since 1926, and corresponding theoretical erythemal UV-Biometer radiation at midday on the 15th (dotted line). Assumption: no snow, constant aerosol conditions.](image)

**Figure 5.8:** Monthly average total ozone amount (solid line) at Arosa during March (above) and June (below) since 1926, and corresponding theoretical erythemal UV-Biometer radiation at midday on the 15th (dotted line). Assumption: no snow, constant aerosol conditions.

The estimations are valid for clear-sky conditions and thus assumed that all atmospheric conditions remained constant from 1970 to 1997. Note that, no significant trend in the aerosol turbidity was observed by comparing 1909 - 1929 to 1930 - 1968 at Davos (Hoyt and Fröhlich, 1983). No trend have been either observed since the 70s, but only increases in the turbidity due to volcanic eruptions which leveled off slowly to the standard value (e.g. El Chichon in 1982, see Heimo and Fröhlich, 1989). An increase in the aerosol content is then not expected to balance the decrease in ozone at Davos. Due to the high cloudiness variability, we cannot, however, claim the existence of a significant trend in the UV radiation received at ground since 1970.
5.4  Normalization to fixed total ozone content and Sun-Earth distance

In this section, we introduce a normalization procedure which filters out the effect of variations in astronomical parameters and in total ozone for the estimation of the influence of aerosol, surface albedo and clouds.

A correction for the Sun-Earth distance is straightforward (see below). The influence of variations in ozone content is more delicate. Schafer et al. (1996) presented an ozone correction based on radiative transfer calculations and dependent on the zenith angle. Some authors used a normalization procedure based on an estimated mean RAF value. However, this method has in our opinion too many uncertainties, since (1) RAF depends on the ozone amount and zenith angle, and (2) variations in radiation and in ozone are not linearly correlated (see section 5.2). We present here a normalization procedure which takes into account the dependence on total ozone and zenith angle, as well as the non-linearity.

5.4.1 Normalization procedure

UV-Biometer measurements $UV_{Bi0}(X, \theta, d_c, \Theta)$ at total ozone $X$, zenith angle $\theta$, Sun-Earth distance correction $d_c$, and other atmospheric properties $\Theta$ are normalized to $UV_{Bi0}(X_0, \theta, d_0, \Theta)$ with following equation

$$UV_{Bi0}(X_0, \theta, d_0, \Theta) = UV_{Bi0}(X, \theta, d_c, \Theta) \cdot CF_{d=1}(d_c) \cdot CF_{X=X_0}(X, \theta)$$

(5.4)

where $CF_{d=1}(d_c)$ is the normalization factor to Sun-Earth distance 1 AU ($d_0 = 1$) and $CF_{X=X_0}(X, \theta)$ is the normalization factor to total ozone amount $X_0$. The equation is valid for global, direct and diffuse UV-Biometer measurements.

The normalization factor to 1 AU Sun-Earth distance

$$CF_{d=1}(d_c) = 1/d_c$$

(5.5)

depends on the Sun-Earth distance correction $d_c$ (see Appendix A). It takes values between 0.966 (minima in winter) and 1.036 (maxima in summer).

We assume that $CF_{X=X_0}(X, \theta)$ depends on total ozone amount $X$ and zenith angle $\theta$, without any influence of the other parameters $\Theta$ (see below). The normalized value $UV_{Bi0}(X_0, \theta, d_0, \Theta)$ depends still on the zenith angle, aerosol, clouds and surface properties.

5.4.2 Normalization to a fixed total ozone amount

The normalization factor $CF_{X=X_0}(X, \theta)$ to a fixed ozone amount $X_0$ is based on the assumption that the ratio between data with different ozone values...
but all other parameters constant (i.e. $\theta$ and $d_c$) is independent on $\Theta$

$$\frac{UV_{Bio\ global}(X_0, \theta, d_c, \Theta)}{UV_{Bio\ global}(X, \theta, d_c, \Theta)} = \frac{UV_{fit}(X_0, \theta, d_c)}{UV_{fit}(X, \theta, d_c)}$$

(5.6)

where $UV_{fit}(X, \theta, d_c)$ is a clear-sky global erythemal reference at total ozone $X$, zenith angle $\theta$ and Sun-Earth distance correction $d_c$. The normalization factor for global erythemal UV-Biometer radiation is then

$$CF_{X=X_0}(X, \theta) = \frac{UV_{fit}(X_0, \theta, d_c)}{UV_{fit}(X, \theta, d_c)}$$

(5.7)

If we assume that the ratio between direct and global radiation is independent of the ozone content, we have the same normalization factor $CF_{X=X_0}(X, \theta)$ for the direct, diffuse and global radiation.

The clear-sky models for Davos (no-snow or snow model) are used as clear-sky reference for the normalization of the UV-Biometer measurements at Davos. This procedure is then valid for total ozone and zenith angle ranges in Table 5.1. The normalization factor $CF_{X=300}(X, d_c)$ for no-snow conditions ranges between 0.74 and 1.11 and equals 1 at 300 DU (Fig. 5.9). The standard deviation is smaller than 1% of the normalization factor.

![Figure 5.9: Normalization factor to 300 DU total ozone as a function of the zenith angle and total ozone amount. No-snow conditions, and Sun-Earth distance 1 AU.](image)

The normalized UV-Biometer measurements for February 23 and 24, 1996 are shown on Fig. 5.10. These days have very different ozone amounts but similar aerosol turbidity (0.1 aerosol optical depth at 368 nm). The normalized data are indeed almost equal at same zenith angle.
5.4. NORMALIZATION TO A FIXED OZONE CONTENT AND 1AU

FIGURE 5.10: Clear-sky global (solid curve), direct (dotted curve) and diffuse (dashed) UV-Biometer data on February 23 (420 DU) and 24 (328 DU), 1996 at Davos. Without (above) and with (below, zenith angle smaller than 70°) normalization to 300 DU and Sun-Earth distance 1 AU. Vertical dotted lines at 56.5° solar zenith angle.

Note that the normalization procedure may theoretically slightly bias the data. The main assumption is that the influence of ozone and the other effects Θ can be separated. In essence, this assumption means that only the intensity below the ozone layer is relevant for the troposphere. Thus, variations in ozone profile are for instance not taken into account. However, no significant bias could be identified in the various tests realized with the data, probably because of the low tropospheric ozone amounts at Davos. Note that the normalization procedure is valid for cloudy skies with the same assumptions as in clear-sky conditions. If not specified, the term "normalized data" means in the following: UV-Biometer measurement normalized to total ozone amount X = 300 DU and Sun-Earth distance 1 AU.

Chapter 5 Summary

Solar zenith angle and total ozone amount allow determining the global UV-Biometer irradiance at Davos with a good accuracy on clear-sky days with a
surface covered by snow or not. Due to larger amount and variation in the aerosol turbidity, such a model is less accurate at Payerne than at Davos.

The global UV-Biometer irradiance on days with a snow free surface is at 30° 11.5 larger than at 70° zenith angle with a total ozone amount equal to 300 DU. A variation in total ozone from 240 to 360 DU leads to a increase by a factor 1.4 at 30° zenith angle. No significant trends in the erythemal radiation have been observed since the 70s in mid-latitudes regions with high density of population despite ozone depletion.

A normalization procedure based on clear-sky models allows filtering the influence of variations in total ozone amount and Sun-Earth distance. The procedure is valid for the global, direct and diffuse UV-Biometer at Davos.
Chapter 6

Aerosol particles and altitude

In this chapter, we present results about the influence of the atmospheric aerosol on erythemal UV radiation on clear-sky and snow free days. The effect of variations in aerosol optical properties is first introduced with UV-Biometer measurements at Davos and Payerne. In a next step, radiative transfer calculations allow estimating the influence of aerosol optical properties on direct, diffuse and global erythemal irradiance. We further investigate the complex effect of the aerosol on the radiation with selected case studies.

Aerosol particle amount and their optical properties are important factors varying the radiation with altitude. The increase of erythemal radiation with altitude is then presented at the end of the chapter. Aerosol impact on radiation on days with a snow covered surface will be treated in chapter 7.

6.1 Measure of turbidity

Large variations in aerosol optical properties lead to rather small variations in global erythemal irradiance. As an example, Koepke et al. (1996) estimated with radiative transfer calculations a relative increase of about 5% in erythemal irradiance by decreasing aerosol optical depth at 550 nm from 0.25 to 0.05. They had a decrease of 5% for a doubling of aerosol optical depths from 0.25 to 0.5. The aerosol extinction is mainly due to scattering, thus redistributing direct into diffuse radiation (see section 3.4).

We have measurements of the direct, diffuse and global parts of erythemal irradiance at Davos and Payerne. The ratio direct/global UV-Biometer radiation is a measure of the aerosol turbidity and can be assumed independent on the ozone amount (see remarks in section 5.4). Note that this ratio should be compared for similar zenith angles and surface albedo.

The aerosol turbidity decreased from October 22 to October 24, 1995 at Davos (Fig. 6.1). The part of direct radiation at 58° zenith angle amounts 1

1In this chapter, direct UV-Biometer values are projected on a horizontal surface.
33% on October 22, 38% on October 23, and 42% on October 24, 1995. However, global UV-Biometer data on October 22 and 23 (same ozone amount) didn’t show any significant differences at given zenith angle.

![UV-Biometer measurements](image)

**Figure 6.1:** Global (solid curve), direct (dotted curve) and diffuse (dashed curve) UV-Biometer measurements on October 22 (left, 257 DU), 23 (middle, 257 DU) and 24 (right, 262 DU), 1995 at Davos. Vertical dotted lines at 58° zenith angle.

### 6.2 Turbidity and UV-Biometer data

Two approaches are used to describe the impact of variations in the aerosol properties on UV-Biometer measurements. We compare the ratio direct/global erythemal irradiance at Davos and Payerne and explain part of the variability observed at Payerne using meteorological parameters.

#### 6.2.1 Variability of the turbidity

The ratio direct/global UV-Biometer radiation is generally larger at Davos (Fig. 6.2) than at Payerne (Fig. 6.3), because of scattering of light by air molecules and aerosol particles in the layer between the altitudes of Davos and Payerne. The ratio direct/global has a larger variability at Payerne than Davos, as the optical properties of the layer between both stations is mainly determined by highly variable natural and anthropogenic particles contents.

The aerosol turbidity at Davos was larger (smaller ratio direct/global) in July - August 1995 than in June, September and October 1995. October 1995 had many low turbidity days (see data between 55 and 70° with large ratio on Fig 6.2). The aerosol turbidity at Payerne was usually larger in
summer than in winter. The mean values were similar in March - April 1996 and September - October 1995, but the variability larger in March - April (Fig. 6.3). The turbidity in October 1995 was not particularly low in Payerne as it was in Davos.

Turbidity varies from year to year, from place to place and is altered by local effects like vicinity of cities or other sources. The annual course at Davos, however, is close to the observed long-term average (Hoyt and Fröhlich, 1983) - except maybe October - and should be representative for stations at similar altitudes in the Swiss Alps. The aerosol conditions at Payerne should also be generally representative for the Swiss plateau as noted by Schmid et al. (1997) (measurements on summer days).
6.2.2 Meteorological parameters to describe turbidity

The deviations between the clear-sky model and the measurements at Pay-erne depended on the ratio direct/global (see section 5.1 and Röösli, 1997) and thus on turbidity. We tried to describe the remaining variability of the UV-Biometer data with meteorological variables such as temperature, wind, and humidity conditions, as well as precipitation in the previous days. The standard deviation of the residuals decreased to 6.5% from 8% for the model without meteorology, the average of the absolute residuals to 5.1% from 6.4%, and the dependence of the deviation on the ratio direct/global diminished. However, the test for another period failed. Thus the addition of meteorological variables did not improve the results significantly. On another hand, a model was fitted with zenith angle, total ozone amount and visibility as input. The error for the analysis data set was similar to that of the control data set (average absolute error around 5.5%).

The use of meteorological parameters in an empirical model could probably improve its accuracy. Such a model has, however, to be fitted with data over many years to get representative correlations.

6.3 Radiative transfer in the aerosol

We performed sensitivity calculations with the radiative transfer model TUV (see chapter 4) before comparing radiative transfer results and measurements at Davos.

6.3.1 Modelling of the aerosol properties

The aerosol optical depth $\tau_a$ is commonly parameterized by the Ångström formula

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

(6.1)

where $\beta$ is the Ångström turbidity coefficient, $\alpha$ is the wavelength coefficient and $\lambda$ is the wavelength [\mu m] (Ångström, 1961 and 1964). The coefficient $\beta$ corresponds to the optical depth at 1 \mu m and varies from 0 to 0.5 or higher. The coefficient $\alpha$ is related to the size distribution of the aerosol particles, and varies from 4 (particles in the order of air molecules) to 0 (very large particles) (Iqbal, 1983).

The Ångström formula is based on the fact that Mie scattering depends on the ratio of the particle radius over the wavelength. Assuming a Junge distribution of the aerosol particles as

$$dN = \frac{k}{r^\nu} dr$$

(6.2)
where \(dN\) is the number of particles within the interval \(dr\) of the radius \(r\), \(k\) is a scaling factor and \(\nu\) the power law exponent of the aerosol distribution, one can show that the extinction follows also a power law, but with exponent \(\alpha = \nu - 3\).

Sensitivity calculations have been conducted with TUV for various wavelength dependencies \((\lambda^{-0.5}, \lambda^{-1}\) and \(\lambda^{-2})\). The results show a significant difference only for very turbid atmospheres and for the direct part of the erythemal radiation (about 3\% larger with \(\alpha = 0.5\) and 4 - 5\% smaller with \(\alpha = 2\) compared to \(\alpha = 1\); \(\tau_e(340\text{ nm}) = 0.8\)). This result is mainly explained by the fact that the aerosol particles with a low exponent have to compete with the strong wavelength dependence of the molecular scattering.

Sensitivity to variations in the profile pattern was analysed by comparing the so-called Denver and non-Denver cases at the altitude of Davos. In the Denver case, the Elterman profile was shifted to get the lower level at the altitude of Davos. In the non-Denver case, the Elterman profile remained with lower level at sea-level (same total aerosol optical depth \(\tau_0(340\text{ nm}) = 0.5\) at the altitude of Davos). The difference between Denver and non-Denver case was smaller than 3\% for zenith angles smaller than 70\°.

In consequence of the above results, and if not specified, we use in the following the Elterman profile without shifting (non-Denver case) and we assume that the aerosol optical depth \(\tau_0\) scales inversely with first power of wavelength (\(\alpha = 1\)). In most cases we don’t have any information on these parameters but their influence is in any case small.

Assuming that the particles optical properties and size distribution do not depend on altitude (i.e. rough approximation), the aerosol influence is determined by three parameters (see definitions in chapter 2): asymmetry factor \(g\), single scattering albedo \(\omega_0\) and total aerosol optical depth \(\tau_a\).

The aerosol optical properties at 300 nm are assumed very similar for typical clean continental, average continental and urban aerosol types. The single scattering albedo \(\omega_0\) ranges between 0.9 and 1, and the asymmetry factor \(g\) between 0.65 and 0.7 (d’Almeida et al., 1991). Shettle (1989) gave asymmetry factors between 0.67 and 0.8 for the urban, rural and tropospheric models at 300 nm. Single scattering albedo and asymmetry factor decrease slightly with increasing relative humidity.

In comparisons between measurements and transfer calculations, we use the parameters at 340 nm and fit direct, diffuse and global irradiances simultaneously. Note that direct erythemal radiation is strongly influenced by variations in total aerosol optical depth \(\tau_a\), and diffuse radiation is mainly influenced by variations of scattering optical depth \(\omega_0\tau_a\). Variations in the asymmetry factor influences mainly the distribution of the diffuse radiation and the amount of radiation backscattered to space.
6.3.2 Extreme aerosol optical properties at Davos

The effect of variations in aerosol properties at Davos can be studied after normalization of the data to 300 DU ozone amount and 1 AU. The normalization procedure described in section 5.4 is applied to the direct, diffuse and global clear-sky UV-Biometer data from May 1995 to December 1996. Only data on days without snow at Davos and Weissfluhjoch (see chapter 5) are considered here.

In the plot of normalized direct and diffuse UV-Biometer data as a function of zenith angle (Fig. 6.4), we note that data with minimal direct radiation (thick curve) correspond to data with maximal diffuse radiation (and vice versa, thin curve). We have then extremes of low and high aerosol loading.

The direct erythemal irradiance is 35 to 65% smaller for very turbid than for very clear conditions (solar zenith angles between 25 and 70°). The diffuse irradiance is 44 to 15% larger in the same range of zenith angle. We have then a decrease in global erythemal radiation equal to 4 to 5.5% when comparing very turbid days to very clear days at Davos. This rather small difference confirm that aerosol extinction is mainly due to scattering (low absorbing aerosol particles and strongly forward scattering functions).

TUV calculations have been conducted to find aerosol optical properties $g$, $\omega_0$ and $\tau_a(340\text{nm})$ which fit the direct and diffuse UV-Biometer data for these two extremes of turbidity.

Very turbid conditions are in good accordance (Fig. 6.4, thick curve) with calculations with total aerosol optical depth $\tau_a(340\text{nm}) = 0.5$, and standard single scattering albedo and asymmetry factor ($\omega_0 = 0.99$, $g = 0.7$). Direct radiation is slightly smaller and diffuse slightly larger with DISORT method than with the two-stream method, due to differences in the treatment of scattering.

In very clear conditions, it wasn’t possible to fit the data with standard $g$ and $\omega_0$, even if we accept a small overestimation of diffuse radiation due to the orography. Two combinations have been however extracted (1) $\omega_0 = 0$, $g = 0.7$, $\tau_a(340\text{nm}) = 0.03$, and (2) $\omega_0 = 0.3$, $g = 0.5$, $\tau_a(340\text{nm}) = 0.03$. As the aerosol optical depth is very small, two-stream and DISORT methods didn’t show significant differences. These combinations correspond to very few and highly absorbing particles. None of them correspond to realistic aerosol properties.

Two explanations for this fitting problem are considered. First, one complementary absorbing gas may miss in the radiative transfer calculations. Second, we may have a problem of accuracy in the model and/or in the data.

Within the gases neglected in the calculations, only SO$_2$ and NO$_2$ absorb UV radiation. Both gases are measured at Davos (NABEL station, BUWAL, 1997). The SO$_2$ ground concentrations in Davos are small and cannot influence significantly the UV radiation (maximum monthly mean = 2.6 $\mu$g/m$^3$.
Figure 6.4: Normalized direct (left), diffuse (middle) and global (right) UV-Biometer data (points) at Davos. Low turbidity (thin curve) and high turbidity (thick curve) model calculations with Two-stream method. CIE (solid curve) and 0.5 UVBio(CH) (dashed curve) weighted irradiance.

on February 1996; maxima around 1 \(\mu g/m^3\) on months without snow). The \(NO_2\) ground concentrations are larger (mean equals 7.6 \(\mu g/m^3\) on February 1996, with maxima at 15 \(\mu g/m^3\)), and the maximum \(NO_2\) column above Jungfraujoch is about 6 \(10^{15}\) molecules/cm\(^2\) (Zander et al., 1996). Such concentrations are equivalent to \(NO_2\) columns smaller than 2 DU, when considering 15 \(\mu g/m^3\) in the first km above Davos and 6 \(10^{15}\) molecules/cm\(^2\) above, thus have no significant influence on erythemal radiation received at ground. As no problem has been detected in the measurement of diffuse radiation at Davos, the fitting problem on very clear days is then probably due to unknown accuracy problems in the radiative transfer calculations.

Note also that we have larger differences between CIE and 0.5 UVBio(CH) weighting irradiances in diffuse than in direct radiation (solid and dashed curves in Fig. 6.4). This is due to the fact that the diffuse part of erythemal radiation contains relatively more shorter wavelengths than direct erythemal radiation and that the difference of sensitivity in the shorter wavelengths mainly determines differences in the weighting irradiances.
6.4 Case studies

Aerosol turbidity was much higher on October 22 than 24, 1995 (Fig. 6.1). We present here more details to understand the differences between these two days (variation in the aerosol particles amount). Complementary remarks are also presented with measurements on July 22 and 26, 1996 (variation in the aerosol size distribution).

6.4.1 October 22 and 24, 1995: measurements

Filter radiometer measurements indicate strongly different aerosol optical properties on October 22 and 24, 1995 (Fig. 6.5). In consequence, the ratio direct/global UV-Biometer data is strongly varying (Fig. 6.6, left). However, global UV-Biometer radiation was almost equivalent despite small differences in ozone amount (Fig. 6.6, right). Note that we have on both days a larger turbidity during the afternoon than during the morning, which explains the asymmetry observed in Fig. 6.6.

![Figure 6.5: Aerosol optical depth at 368 nm, 412 nm, 500 nm and 778 nm (from top to bottom) on October 22 (left) and 24 (right), 1995 at Davos. Dotted vertical curve at solar noon.]

The Ångström parameterization of the aerosol optical depth (Eq. 6.1) can be used to get information about the aerosol particles. At each time with simultaneous \( \tau_a \) values at various wavelengths, the Ångström coefficients \( \alpha \) and \( \beta \) can be estimated with a regression model:

\[
\log(\tau_a(\lambda)) = \log \beta - \alpha \log(\frac{\lambda}{1000}) + \epsilon
\]

where \( \lambda \) in (368 nm, 412 nm, 500 nm, 778 nm) and \( \epsilon \) is the error. Note
that only undisturbed data (without cloud influence) should be used for the estimation of $\alpha$ and $\beta$.

At midday, the wavelength coefficient $\alpha$ was similar on 22 and 24 (Fig. 6.7), which means a similar aerosol particles size distribution, corresponding to typical extra eruption $\alpha$ values at Davos (Schmid et al., 1997). Only the amount of particles, represented by $\beta$, is a factor of 4 - 6 larger on October 22 than 24. Note that the Ångström parameterization is less accurate on October 24 than 22 because of low turbidity (larger standard deviation and variability of $\alpha$, see Table 6.1).

The large difference in aerosol particles amount may be explained by the
meteorological situation. On October 22, we have over Davos an air mass coming from the north-east of Europe. A high pressure system grew over the centre of Europe and no precipitation have been measured at Davos since October 1. We have then an accumulation of aged aerosol particles (accumulation range, 0.1 - 1 \( \mu m \) radius). Wind measurements at Davos didn't show clear difference between October 22 and 24 due to the dominant valley wind system (only the change from S-W to N-E around midday is shifted). The synoptic situation, however, changed between October 22 and 24 (Fig. 6.8 and Table 6.1). The air mass over Davos on October 24 came from eastern of Europe via Italy. It had therefore a completely other history then the air mass on October 22.

\[ \text{Figure 6.8: 72 hours backward trajectories at 800 hPa, 750 hPa, ..., 150 hPa on October 22 (left) and 24 (right), 1995 12 UTC. Calculations from 5 grid points around Davos based on ECMWF analysis data (H. Wernli, LAPETH).} \]

\[ \text{Table 6.1: Optical and meteorological parameters around midday on October 22 and 24, 1995 at Davos. Ratio direct/global UV-Biometer at } \theta = 58^\circ. \text{ Stratus over north-east part of Switzerland on October 24. Note that the visibility is mainly determined by particles in the accumulation range.} \]

<table>
<thead>
<tr>
<th></th>
<th>October 22</th>
<th>October 24</th>
</tr>
</thead>
<tbody>
<tr>
<td>Direct/global UV-Biometer</td>
<td>33%</td>
<td>42%</td>
</tr>
<tr>
<td>Angström parameters</td>
<td></td>
<td></td>
</tr>
<tr>
<td>( \alpha = 1.55 \pm 0.03 )</td>
<td>( \alpha = 1.55 \pm 0.3 )</td>
<td></td>
</tr>
<tr>
<td>( \beta = \sim 0.04 )</td>
<td>( \beta = \sim 0.003 )</td>
<td></td>
</tr>
<tr>
<td>Ozone amount</td>
<td>257 DU</td>
<td>262 DU</td>
</tr>
<tr>
<td>Visibility</td>
<td>60 km</td>
<td>&gt; 70 km</td>
</tr>
<tr>
<td>Surface albedo</td>
<td>no snow</td>
<td>no snow</td>
</tr>
<tr>
<td>Dominant wind direction</td>
<td>N-N-E</td>
<td>S-W</td>
</tr>
<tr>
<td>Wind velocity</td>
<td>2 - 3 m/s</td>
<td>1 - 3 m/s</td>
</tr>
<tr>
<td>Temperature</td>
<td>11°C</td>
<td>24°C</td>
</tr>
<tr>
<td>Relative humidity</td>
<td>42%</td>
<td>13%</td>
</tr>
</tbody>
</table>
6.4.2 October 22 and 24, 1995: modelling

On October 22 and 24, 1995, we have information about total ozone amount at Arosa and aerosol optical depth at Davos. The optical depth $\tau(340\text{nm})$ and the wavelength dependence $\alpha$ can be derived by using Ångström formula. Radiative transfer calculations were then performed for comparison with UV-Biometer measurements.

With asymmetry factor $g = 0.7$, single scattering albedo $\omega_0 = 0.99$ and surface albedo 0.03, direct radiation is well approximated by transfer calculations. However, diffuse radiation is overestimated by 10 and 20% with larger deviation on 22 than 24, i.e. larger for larger turbidity (Fig. 6.9 and 6.10). This overestimation is larger than expected from an influence of the orography around the station (see section 4.6). Aerosol properties combination $g = 0.7$ and $\omega_0 = 0.8$ allow better fitting the data on October 22. On October 24, the diffuse radiation is still slightly overestimated with the unrealistic value for Davos $\omega_0 = 0.7$. Once again (see 6.3.2), $\omega_0$ values lower than expected would be required to get a good fit between calculations and measurements of diffuse irradiance.

![Figure 6.9: Global, direct and diffuse UV-Biometer measurements (points) on October 22, 1995 at Davos. DISORT calculations with CIE (solid line) and 0.5 UVBio(CH) (dashed curve) weighted irradiance.](image)

6.4.3 July 22 and 26, 1996

An interesting case is also found on July 22 and 26, 1996 (data until 13 UTC). The ratio direct/global is similar on 22 and 26 (Tab. 6.2), but the patterns of the aerosol optical depths are very different (Fig. 6.11).

The Ångström parameters vary strongly between the two days (Fig. 6.12). The larger amount of particles on 26 (larger $\beta$ parameter) compensated the presence of much larger particles (smaller $\alpha$ parameter). As a result, the
aerosol optical depth at 368 nm remains almost constant. The interval between aerosol optical depths (groupings of wavelengths) is changing from July 22 to July 26. This pattern gives information on the aerosol size distribution. Actually, the parameterization of Ångström is less adequate on 26 than 22 (worst fit) (standard deviation of $\alpha$ on Fig. 6.12). The size distribution is then probably not well approximated by Junge's power law, it may, for instance, be a two modes distribution. The variation in the aerosol properties is probably due to the precipitation on 23 to 25. On 22 we have aged aerosol particles (accumulation range) and on 26 fresh ones (possibly influence of nearby source).

Note that the remarks above are based on calculations with only four wavelengths. Aerosol optical depth at complementary wavelengths might give more information on the aerosol properties. The Ångström formula is a parameterization to get general information on the amount and dominant size particles, and any too detailed interpretation may not be reliable. See
6.5 Altitude effect

The increase of global solar irradiance with altitude is usually given as a percentage increase over 1000 m relative to the lower station with clear-sky conditions. Relative increases are found between about 5 and 24% /1000 m (erythemal and UV-B; summary in Weihs et al., 1995). Blumthaler (1993) found an increase of 18 ± 5%/1000 m in summer, and 23 ± 6%/1000 m in winter by comparing Robertson-Berger data at Innsbruck (577 m a.s.l.) and at Jungfraujoch (3576 m a.s.l.). The estimation was based on daily doses under undisturbed conditions. The altitude effect between annual doses of UV-B radiation at Jungfraujoch and Innsbruck corresponded to 19%/1000 m (Blumthaler et al., 1992).

Koepke et al. (1996) calculated a relative difference around 11 - 15% between erythemal irradiance at 1500 m a.s.l. and 0 m a.s.l. The value ranged between 17 and 22% between 2500 m a.s.l. and 0 m a.s.l. The increases per 1000 m would then be about 7.3 - 10%/1000 m, and 6.8 - 8.8%/1000 m respectively. The relative difference increases for increasing solar zenith angle up to 60 - 70° and decreases for larger zenith angles.

Measurements at high altitude stations are larger than at low altitude stations, with a larger part of direct radiation (Fig. 6.13). The 1120 m thick layer between Davos and Payerne contains aerosol particles, as well as
scattering and absorbing gases. Variations in the properties of this layer and local effects lead to variations in the altitude difference. Furthermore, the presence of clouds (stratus between both stations or different cloud coverage) may obviously lead to large variations between stations.

Figure 6.12: Ångström parameters $\alpha \pm$ standard deviation (left) and $\beta$ (right) on July 22 (solid curve) and 26 (dashed curve), 1996 at Davos.

Figure 6.13: Global (solid curve), direct (dotted curve) and diffuse (dashed curve) UV-Biometer measurements at Davos (left) and Payerne (right) on May 22, 1995.
6.5. ALTITUDE EFFECT

6.5.1 Global UV-Biometer and altitude

We present three different estimations of the altitude effect between Davos and Payerne. The first is based on envelopes around all-weather global UV-Biometer data. The second deals with the ratio between global clear-sky measurements, and the third is based on clear-sky models described in chapter 5.

The ratio between functions enveloping the data at Davos and Payerne allow estimating the altitude effect. The envelopes may be determined by hand or by using an adequate automatic algorithm (example for Davos in Fig. 6.14). The data are 18 - 39% larger in Davos than Payerne in summer (May - September) by using hand made envelopes. Corresponding values equal 17 - 30% with the automatic algorithm proposed in Appendix C ("moving quantile method"). Estimation for winter are in both cases not very reliable.

![Global UV-Biometer Measurements](image)

**Figure 6.14:** Global UV-Biometer measurements (11:30 - 12:30 UTC) at Davos as a function of Julian day with hand made envelope (solid curve) as well as automatic envelope (dashed curve, see algorithm in Appendix C). May 1995 - April 1996. Parameterization of the hand-made curve: \( k_0 = 0.021, k_1 = -0.126, k_2 = 6.08, k_3 = -11.36, k_4 = 0 \), and of the automatic curve: \( k_0 = 0.03, k_1 = -0.076, k_2 = 3.335, k_3 = 0.685, k_4 = -21.457 \) (see formula in Appendix C).

The variability of the ratio between clear-sky measurements at Davos and Payerne is very large (Fig. 6.15), despite the fact that only measurements with sunshine at Payerne and Davos were used. Measurements are on average 20 to 40% larger at Davos than at Payerne (smooth curve) during summer months. Moreover, the altitude effect depends on the solar zenith angle. We have, for instance, data which are 42% larger at Davos than at Payerne on May 22, 1995 at solar noon (zenith angle 26°), and only 23% larger at 70°.
The clear-sky models calibrated with clear-sky data at Davos and Payerne (see chapter 5) allow an estimation of the altitude effect as a function of total ozone and zenith angle. Data in Davos are found 14 to 30% larger than in Payerne with average 22%.

Note that the estimations above are based on measurements during one year and could vary over the years. Furthermore, variability due to different surface albedo is not specifically taken in account. However, these values give information on the difference between Davos and Payerne (around 20 - 40%) and its large variability. The 20 - 40% relative difference between Payerne and Davos corresponds to about 18 - 36%/1000 m with reference to Payerne. On May 22, 1995 the increase around midday was 37.5%/1000 m.

We have then altitude effects per 1000 m which are in general larger than the estimates given in Weih et al. (1995), Blumthaler (1993) and Blumthaler et al. (1992), and a significant zenith angle dependence (smaller values for larger zenith angles).

6.5.2 Optical depth between Davos and Payerne

The average optical depth (weighted with erythemal response) of the layer between Davos and Payerne can be approximated from Beer-Bouguer-Lambert's law

\[ r_{Dav-Pay} = - \frac{1}{m} \log \left( \frac{UV_{Bio\,direct}^{Pay}(X, \theta)}{UV_{Bio\,direct}^{Dav}(X, \theta)} \right) \]  

(6.4)

where \( UV_{Bio\,direct}^{Pay}(X, \theta) \) and \( UV_{Bio\,direct}^{Dav}(X, \theta) \) are direct UV-Biometer measurements at Payerne and Davos, \( X \) is total ozone amount at Arosa and \( \theta \) is the solar zenith angle. Assuming a plane-parallel homogeneous layer, we have \( m = 1/\cos \theta \).
The layer optical depth $\tau^{\text{Dav-Pay}}$ is the sum of Rayleigh, aerosol and ozone optical depths and varies strongly with time (Fig. 6.16) due to variations in tropospheric ozone and aerosol content. The Rayleigh optical depth $\tau^{\text{av}}_{\text{Pay}}$ can be estimated with

$$
\tau^{\text{Dav-Pay}}(\lambda) = \frac{\tau_r(\lambda)}{p_0}(p^{\text{Pay}} - p^{\text{Dav}})
$$

(6.5)

where $\tau_r(\lambda)$ is the total Rayleigh optical depth at sea level (Appendix B), $p_0 = 1013.25$ hPa is the pressure at sea-level, and $p^{\text{Pay}}$ and $p^{\text{Dav}}$ are the pressure values at Payerne and Davos. Since the optical depth $\tau^{\text{Dav-Pay}}(\lambda)$ equals 0.15 at 290 nm, 0.1 at 320 nm and 0.04 at 400 nm with mean pressure $p^{\text{Pay}} = 950$ hPa and $p^{\text{Dav}} = 840$ hPa, the sum of the ozone and aerosol optical depths between Davos and Payerne is about $\tau^{\text{Dav-Pay}} - 0.1$.

As expected, the Rayleigh optical depth is thus much smaller than the combined aerosol and ozone optical depth. We note that the sum of the aerosol and ozone optical depth between Davos and Payerne is often larger than that of the total aerosol optical depth over Davos (smaller than 0.5 at 340 nm, see above).

On May 22, 1995, $\tau^{\text{Dav-Pay}}$ equalled 0.5 around midday and decreased for increasing zenith angles to 0.3 at 60°. This dependence on the zenith angle indicates that assumptions in the calculation of $\tau^{\text{Dav-Pay}}$ are not completely fulfilled. However, equation 6.4 should give a good approximation for the real layer optical depth, as time variations are larger than the zenith angle dependence.
6.5.3 Comparison of altitude effects

The parameter "relative increase per 1000 m" should be used carefully. Actually, the expected exponential behaviour of attenuation with altitude may be approximated by a linear relation only for small optical depths and zenith angles. This parameter varies with varying reference station and zenith angle.

First, the increase of $z_1\%/1000\ m$ with reference station 1 at altitude $h_1$ is related to the $z_2\%/1000\ m$ increase with reference station 2 at altitude $h_2$ by the following formula (same absolute increase)

$$x_2 = \frac{1000}{(h_2 - h_1)} \left[ 1 + \frac{100}{x_1} \frac{1000}{(h_2 - h_1)} \right]^{-1}$$

(6.6)

An increase of 22%/1000 m with reference to Payerne corresponds then to 18%/1000 m with reference to Davos.

Second, the relative increase $x\%/1000\ m$ varies with varying zenith angle, even if the 1000 m thick layer has a constant optical depth $r_{layer}$. The dependence

$$x = (e^{-r_{layer}m} - 1) \times 100$$

(6.7)

is valid for the direct and in first approximation also for the global irradiance, where $m = 1/\cos \theta$ and $\theta$ is the solar zenith angle.

Estimations of increases per 1000 m should only be compared for equivalent zenith angle and same reference altitude. Eq. 6.6 can be used to adapt values to a given reference altitude (for instance sea level) before comparison. As an example, 18 - 36%/1000 m with reference to Payerne corresponds to 20 - 44%/1000 m with reference to sea level, which can be compared to the adapted value 20%/1000 m from Blumthaler (1993) (with assumption of no zenith angle dependence).

A more adequate parameter would be based on an exponential law to take into account changes due to varying zenith angles (e.g. optical depth of a 1000 m thick layer), or on the scale height (thickness corresponding to an attenuation by factor 1/e). The choice of two linear increases in the boundary layer (higher value) respectively in the "free troposphere" (lower value) should provide more general information on the altitude effect (example of profiles in chapter 3).

Note that calibration problems and local influences (shadowing, mountains, regional albedo) are expected to influence the results. Measurement at Arosa (1820 m a.s.l.) and Davos (1610 m a.s.l.) show for instance that the altitude effect is partially influenced by local effects. Data at Tschuggen (2040 m a.s.l.) were also found smaller than at Arosa (1820 m a.s.l.), despite higher altitude (Wiedemann, 1997; analysis of UVB-1 pyranometer Yankee data).
Chapter 6 Summary

The aerosol turbidity is larger and varies more strongly at Payerne than at Davos. Variations in aerosol properties at Davos lead to maximum 4 - 5.5% variations in the global erythemal radiation.

The aerosol turbidity at Davos ranges between two extremes, which correspond to aerosol optical depths $\tau(340\text{nm})$ with very small values up to 0.5 on clear-sky days with a snow free surface. The radiative transfer model calculations do not fit data on very clear days with realistic aerosol optical properties. This problem seems to be due to an overestimation of the diffuse erythemal irradiance.

The difference of erythemal radiation between Payerne (490 m a.s.l.) and Davos (1610 m a.s.l.) varies strongly during the year and between days. The mean relative difference between Payerne and Davos ranges between 20 and 40%. The aerosol optical depth of the layer between Davos and Payerne (polluted boundary layer) is often larger than the aerosol optical depth over Davos.
Chapter 7

Surface albedo

Results about the impact of the surface albedo on erythemal UV radiation are presented in this chapter. The effect is first estimated on the basis of UV-Biometer measurements. Radiative transfer calculations allow then the estimation of the redistribution into direct and diffuse erythemal radiation for winter aerosol conditions. The non-lambertian behaviour of the surface reflection is also discussed.

Only clear-sky measurements at Davos are considered in this chapter. Results concerning the interaction between cloud and surface are described in chapter 8. Note also that the data are horizontal measurements. The effect of an inclination of the target is not considered (see information about the inclination effect e.g. in Blumthaler et al., 1996).

7.1 Reflection on the ground

Most of the knowledge about the influence of the surface albedo on UV radiation comes from radiative transfer calculations. As an example, Kylling (1993) estimated an increase of downward irradiance by factor 1.8 at 320 nm for an albedo equal to 1 and by factor 1.22 for albedo equal to 0.5, in an sub-arctic summer atmosphere and solar zenith angle 60°. The effect was smaller for shorter as well as for larger wavelengths. Koepke et al. (1996) found an increase relative to an albedo 0.03 of about 8% with an albedo 0.2, 22% with 0.5, and 42% with 0.8 (330 DU ozone, 0 m a.s.l., 0.25 aerosol optical depth at 550 nm). Ambach et al. (1991) also estimated 18% more UV radiation in February - March than in September - October in Fairbanks by using a simple model. Some other measurements and modelling of the snow effect, mainly in the Arctic, can be found in Ambach et al. (1991).

The surface albedo has an influence on the diffuse part of radiation (see section 3.5). The aerosol content also interact with the direct - diffuse redistribution (see section 3.4 and chapter 6). We have then to pay special
attention to the aerosol in the estimation of the albedo effect on diffuse radiation. Furthermore, the influence of the ozone content (stratospheric as well as tropospheric) has also to be taken into account in the various estimations.

The surface was free of snow at Davos on October 24, 1995 and covered by snow on February 28, 1996. Both days have a similar aerosol optical depth (around 0.04 at 368 nm). As a consequence, the part of diffuse radiation in the global radiation is smaller on October 24 (58% at solar zenith angle 58°) than on February 28, 1996 (66% at same zenith angle) (Fig. 7.1).

![Figure 7.1: Global (solid curve), direct (dotted curve) and diffuse (dotted curve) UV-Biometer measurements on October 24, 1995 (left, 262 DU) and February 28, 1996 (right, 358 DU) at Davos. Vertical dotted lines at zenith angle 58°.](image)

### 7.2 Global UV-Biometer and snow

We present two estimations of the increase of global erythemal UV radiation due to a snow covered surface (snow effect) at Davos. The effect was first estimated as a function of the zenith angle after normalization to mean Sun-Earth distance and 300 DU total ozone amount (see normalization procedure in section 5.4). Then the snow and no-snow clear-sky models (see chapter 5) are compared as a function of the zenith angle and total ozone amount.

The measurements are sorted into snow and no-snow data. The criterion is the presence of snow, respectively a snow free surface, simultaneously at Davos and at Weissfluhjoch (2690 m a.s.l., about 6 km from Davos). This criterion was already used in chapter 5 and allows avoiding regional effects on the radiation (see section 3.5).

The clear-sky normalized global UV-Biometer data (May 1995 - December 1996) with a snow covered surface are 17.5 ± 0.5% to 27 ± 1% larger than
data with a snow free surface for zenith angles between 40 and 70° (Fig. 7.2). The value increases with increasing zenith angle. Values for zenith angle larger than 70° are not reliable. The direct normalized UV-Biometer data do not show clear differences between snow covered and snow free surfaces, whereas the diffuse is clearly larger for a snow covered surface. This behaviour confirms what we expect from a variation in the albedo.

Comparison between snow and no-snow models (chapter 5) leads to an increase due to snow which ranges from 10 to 27% (Fig. 7.3, standard deviations smaller than 1% in the considered ozone and zenith angle ranges). The increase at 300 DU total ozone amount equals 15.4% at 40° and 25.3% at 70°, which is very similar to the mean effect estimated above. The snow effect is larger for larger solar zenith angles, and does not depend significantly on the total ozone amount, as tropospheric ozone can be neglected. Note that the comparison between models is valid only for total ozone amounts and zenith angles observed in snow free and snow conditions.

The relative difference between global UV-Biometer data at Davos on days without snow (summer) and days with snow (winter) ranges then between 15 - 17% at 40° and 25 - 27% at 70°. The highly significant solar zenith dependence will be discussed below (non-lambertian behaviour), as well as the effect of variation between winter and summer aerosols.
7.3 Radiative transfer and surface albedo

Comparisons between measurements and radiative transfer calculations with a snow covered surface are more complex than with a snow free surface. Assumptions have to be made about the ground reflection and the mean surface albedo is usually unknown. A measured local albedo (examples in section 3.5) is seldom representative for a mean regional albedo. Some authors tried to extract mean surface albedo values from satellite measurements but the accuracy remains questionable and the directional reflectivity is not taken into account (review in Weihs and Webb, 1997a).

Note that all results based on radiative transfer calculations have to be handled with great caution. Uncertainties remain for large zenith angles as well as high surface albedo. For instance, Weihs and Webb (1997b) presented comparisons between DISORT calculations and measurements at Jungfraujoch at 380 nm, and found a surface albedo equal to 0.17. They had thus good correspondence with an albedo clearly smaller than the real one.

The radiative transfer model TUV assumes that the reflection on ground is lambertian and independent on the wavelength. In comparison of radiative transfer calculations and measurements on days with a snow covered surface, some complementary assumptions are necessary. In this section, we use standard aerosol single scattering albedo $\omega_0 = 0.99$, and asymmetry parameter $g = 0.7$. 

![Figure 7.3: Relative difference between the snow and no-snow models for Davos, as a function of total ozone amount and zenith angle.](image)
7.3.1 Winter aerosol properties

Typical winter optical properties at Davos may be determined by comparing TUV calculations and UV-Biometer measurements. The ratio direct/global UV-Biometer on days with snow varies with varying aerosol properties and surface albedo (Fig. 7.4). As the variability is smaller than on days without snow (summer, see Fig. 6.2), the winter aerosol optical depth may be defined by an average value.

Before comparing calculations and measurements, we normalized the UV-Biometer measurements between May 1995 - December 1996 to 300 DU and 1 AU (normalization procedure, see section 5.4). Only measurements on days with snow at Davos and Weissfluhjoch are considered.

The comparisons between TUV and winter normalized UV-Biometer measurements are performed in two steps. First, the aerosol optical depth $\tau(340 \text{ nm}) = 0.05$ (rather low turbidity days) is estimated by fitting calculated direct erythemal irradiance and direct UV-Biometer data (not influenced by the surface albedo). Second, a surface albedo $r_s$ between 0.3 and 0.5 envelops the diffuse UV-Biometer data. Surface albedo around 0.5 is still valid with a larger aerosol optical depth $\tau(340 \text{ nm}) = 0.1$.

The increase of global UV-Biometer data due to a snow covered surface with $\tau(340 \text{ nm}) = 0.05$ and $r_s = 0.3$ is about 10%. The value is about 19 - 20% with $\tau(340 \text{ nm}) = 0.05$ and $r_s = 0.5$, as well as with $\tau(340 \text{ nm}) = 0.1$ and $r_s = 0.5$ (slightly larger). The snow effect is independent on the zenith angle since the calculation assumes lambertian reflection. The calculated estimates are slightly smaller than the estimate directly based on the UV-Biometer measurements, as we took into account the fact that the winter
aerosol has a lower attenuation than the summer aerosol. Note that global radiation with winter aerosol without snow corresponds to very clear summer days (for same ozone amount and zenith angle).

A surface albedo of 0.3 to 0.5 is rather small for a snow covered surface (see section 3.5). Inhomogeneity in the surface around the station (snow, rock, etc.) could be the reason for a lower albedo than expected for snow. The orography should, however, not perturb the results significantly as shown in section 4.6. The rather low albedo could also be due to an overestimation of the diffuse irradiance in the model (see chapter 6).

Some case studies with data at Davos on days with filter radiometer measurements and a snow covered surface have been analysed in Schmucki (1998). In general, direct irradiance was well fitted by TUV calculations (two-stream method), and diffuse radiation was fitted with surface albedo between 0.35 and 0.5.

### 7.3.2 Surface albedo and atmosphere albedo

The increase of radiation due to a snow covered surface depends on the optical properties of the surface and atmosphere. An atmosphere with high backscattering ability leads to a larger increase of radiation.

The radiation can be expressed as the sum of three components: direct irradiance $\text{dir}$, diffuse irradiance not reflected by the earth's surface $\text{dif(sky)}$, and diffuse radiation which was reflected by the surface and scattered back again to the surface $\text{dif(surf)}$ (Fig. 7.5). With the assumption of a lambertian surface we have (adapted from Liou, 1980)

$$\text{dif(surf)} = \frac{\bar{r} r_s}{1 - \bar{r} r_s} (\text{dir} + \text{dif(sky)})$$  \hspace{1cm} (7.1)

where $r_s$ is the surface albedo and $\bar{r}$ is the portion of radiation backscattered by the entire atmosphere or spherical albedo. The spherical albedo $\bar{r}$ is expected to increase with turbidity (more scattering particles). The factor $\bar{r} r_s/(1 - \bar{r} r_s)$ represents the multiple interactions between the surface and the atmosphere. It depends on wavelength and on the optical properties of the surface and atmosphere. Components $\text{dir}$ and $\text{dif(sky)}$ are independent of the surface albedo. Component $\text{dif(surf)}$ depends on the surface and atmospheric albedos. It is negligible in case of snow free surface with a very small $r_s$ in the UV range, but not in case of a snow covered surface.

For winter aerosol conditions at Davos, i.e. $\tau(340 \text{ nm}) = 0.05$ and $r_s = 0.5$, the calculated diffuse irradiance $\text{dif(surf)}$ equals 16.2% of global erythemal irradiance (Fig. 7.6). With assumptions of lambertian reflection and surface albedo independent of the wavelength, we get a spherical albedo $\bar{r} = 0.324$. If the optical depth is increased to $\tau(340 \text{ nm}) = 0.1$ the spherical albedo increases only slightly to $\bar{r} = 0.328$. 
7.4 Is the surface lambertian?

An important zenith angle dependence is observed for the influence of a snow covered surface on global UV-Biometer data at Davos (see Fig. 7.3).

The aerosol contribute to this zenith angle dependence. Comparisons between calculations (DISORT) for winter with low turbidity ($\tau(340\,nm) = 0.05$, $r_s = 0.5$) and summer with high turbidity ($\tau(340\,nm) = 0.5$, $r_s = 0.03$) show differences from 30% at 40° to 34% at 70°. These atmospheric properties constitute the largest differences we can expect at Davos on clear-sky days. The zenith angle dependence observed in the data is then not completely explained by differences in aerosol properties.

A cosine error in the measurement may be further reason for the remaining
solar zenith angle dependence. We cannot exclude this although the specifications of the UV-Biometer show negligible deviations for solar zenith angles smaller than 70° (see section 4.2).

A wavelength dependence of the surface albedo could also explain part of the zenith angle dependence. Large albedo differences in the UV range are, however, not expected.

It is difficult to estimate the effect of the orography around the station. However, the effect should not be large (see section 4.6). Note also that no dependence on the azimuth has been observed in the data. The snow effect in the morning and afternoon for equivalent zenith angles are not significantly different. This may be explained by the fact that mountains at Davos are not clearly asymmetric and that the received radiation comes from the surrounding region of about 10 km or more.

The observed solar zenith dependence is thus a consequence of a non-lambertian behaviour of surface reflection (inhomogeneities in the terrain). The most probable effect is the high glazing reflectance at low incidence angles. As larger part of radiation is diffuse at large zenith angle, the non-lambertian behaviour of reflection around Davos has probably less influence on the received radiation than at smaller zenith angles.

Chapter 7 Summary

The relative difference between global UV-Biometer data at Davos on days without snow (summer) and days with snow (winter) ranges between 15 - 17% at 40° and 25 - 27% at 70°. The snow effect increases with increasing zenith angle.

By comparing measurements and radiative transfer calculations, we found typical winter aerosol optical depth at Davos \( \tau(340\,\text{nm}) \) between 0.05 and 0.1. The surface albedo fitting diffuse radiation ranges between 0.3 and 0.5. Assuming lambertian reflection, the increase of the calculated global radiation due to snow is about 19 - 20% for typical winter aerosol properties.

The increase of irradiance observed in the measurements at Davos and due to a snow covered surface remains dependent on the zenith angle when extracting variations due to the aerosol. This strongly significant non-lambertian behaviour is probably due to the high glazing reflectance at low incidence angles.
Chapter 8

Clouds

In this chapter, we investigate the impact of clouds on erythemal UV radiation. The attenuation as well as the variability due to clouds are analysed for various cloud cover types.

We compare radiative transfer calculations and measurements to determine the range and variability of an equivalent cloud optical depth at the station Davos (overcast and broken cloud cover). Estimations of the interaction between a high albedo surface and a cloud cover are also part of this chapter. Furthermore, a comparison between UV-Biometer and pyranometer data at Davos leads to information about the wavelength dependence of the cloud impact on radiation.

8.1 Cloud types, observations and radiation

The presence of clouds leads to large variations in erythemal UV irradiance. Each cloud cover has its own way to disturb radiation. It can strongly attenuate as well as enhance the irradiance at ground (see section 3.6). Many studies (measurements and modelling) have been conducted to study the interaction between UV irradiance and clouds.

Bener (1964) investigated the influence of various cloud types at Davos at wavelengths 330 nm and 370 nm and presented results with coverage between 8/10 and 10/10 as a function of the solar zenith angle and season. The overall means of the ratio between the diffuse intensity obtained with and without clouds ranged between 0.7 and 1.1, and grew larger in the average towards smaller solar zenith angle.

Blumthaler et al. (1994) deduced from measurements at Jungfraujoch (3576 m a.s.l.) that erythemal irradiance reduced in average to the 70% level of clear-sky irradiance at 10/10 cloudiness (overcast sky). The cloudiness effect was mostly dependent on the presence of clouds covering the sun (reduction to the 62% level). The relative influence of cloudiness (0/10 - 10/10)
on global erythematic radiation was approximately equal at all solar elevations.

Ambach et al. (1995) found a transmission\(^1\) for daily totals of global erythematic irradiance at Innsbruck (577 m a.s.l.) with a median of 41.8% and a minimum of 9.4%.

Schafer et al. (1996) found average UV-B transmission of 29.6 ± 2.5% at 50° solar zenith angle for overcast skies (10 tenths) at Black Mountain (951 m a.s.l.). The mean transmission was 61.4 ± 8% for 8 - 9 tenths cloud cover, 74.7 ± 7.5% for 6 - 7 tenths and 78.8 ± 7.6% for 4 - 5 tenths. The transmission increased by approximately 17% between 30 and 60° when the cloud coverage was smaller then 10 tenths; no zenith angle dependence was observed for 10 tenths, as the radiance is almost homogeneous distributed.

Many calculations with more or less complicated radiative transfer models in clouds have been performed. As an example, Koepke et al. (1997) estimated cloud attenuation factors for various cloud optical depths. Values over grass were approximately 0.65 with cloud optical depth \(\tau_{ol} = 8.3\), 0.5 with \(\tau_{ol} = 16.6\), and 0.35 with \(\tau_{ol} = 33.2\) (solar zenith angle = 45°). Values over a snow covered surface were around 0.8 with cloud optical depth \(\tau_{ol} = 16.6\). Madronich (1993) also noticed the large impact of the ground albedo on the ground irradiance below a cloud layer.

Note also that some authors developed empirical models to describe UV irradiance under cloudy skies. Frederick and Steele (1995) explained 61 - 77% of the variability in irradiance (total shortwave and erythematic) with 3 parameters (fractional cloudiness, cloud-ceiling altitude, visibility). Bodeker and McKenzie (1996) inferred UV levels with and without clouds from meteorological data, global shortwave measurement and column ozone amounts using a semi-empirical model. Burrows (1997) used a tree-based regression model with 8 predictors (total opacity, liquid precipitation, etc.) to predict UV radiation in the presence of clouds.

The thickness of a cloud cover is generally smaller in the high mountains; thus reductions due to cloudiness have been proven to be greater in low altitude stations. Measurements and modelling show the high variability of attenuation, the presence of a zenith angle dependence and the importance of the interaction with the surface albedo. All these points will be analysed in the following with UV-Biometer measurements and radiative transfer calculations.

At Payerne, we have UV-Biometer measurements as well as observations of the cloud cover and cloud type every 3 hours (see section 4.4). Some examples of measurements are given in Fig. 8.1 with the corresponding observations in Table 8.1. UV-Biometer measurements vary strongly with stratocumulus

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\(^1\)Note that transmission means commonly the ratio “global irradiance with clouds/global irradiance without clouds” in comparisons between clear-sky and cloudy-sky measurements (also called cloud attenuation factor). It should not be mixed up with the transmissivity defined in section 2.4.
8.2. SUNSHINE DURATION AND CLOUD COVERAGE

clouds (see e.g. July 23, 1995). Direct radiation disappears and diffuse radiation is reduced under a thick cloud cover (see e.g. July 24, 1995 during the morning). Direct radiation varies only slightly and diffuse radiation remains rather smooth in the presence of high clouds (see e.g. November 9, 1995). Direct radiation is from time to time strongly reduced by scattered low clouds (sun - cloud - station geometry; see e.g. cumuli on August 23, 1996).

**Figure 8.1:** UV-Biometer measurements at Payerne on July 23 and 24, 1995, November 9, 1995, and August 23, 1996. Global (solid curve), direct (dotted curve), and diffuse irradiance (dashed curve). Vertical dotted lines at 6, 9, 12, 15 and 18 UTC (time of observations).

8.2 Sunshine duration and cloud coverage

The sunshine duration measurements are used in many studies as main criterion to distinguish between the clear and overcast skies. This criterion has also been used (combined with others) in the previous chapters (clear-sky data extraction).

A comparison between sunshine duration measurements and synoptic observations at Payerne leads to some remarks about the relation between these
Table 8.1: Synoptic clouds observations at Payerne. Sc: stratocumulus, Cu: cumulus, Ac: altocumulus, Ci: cirrus. Two layers on July 23, 1995 at 18 UTC.

<table>
<thead>
<tr>
<th></th>
<th>9 UTC</th>
<th>12 UTC</th>
<th>15 UTC</th>
<th>18 UTC</th>
</tr>
</thead>
<tbody>
<tr>
<td>July 23, 1995</td>
<td>6/8 Sc</td>
<td>6/8 Sc</td>
<td>7/8 Sc</td>
<td>2/8 + 5/8 Sc Sc</td>
</tr>
<tr>
<td>July 24, 1995</td>
<td>7/8 Sc</td>
<td>7/8 Sc</td>
<td>1/8 Sc</td>
<td>cloudless</td>
</tr>
<tr>
<td>Nov. 9, 1995</td>
<td>3/8 Ci</td>
<td>4/8 Ci</td>
<td>5/8 Ci</td>
<td>6/8 Ci</td>
</tr>
<tr>
<td>Aug. 23, 1996</td>
<td>1/8 Sc + 3/8 Cu</td>
<td>3/8 Cu</td>
<td>1/8 Cu ext.</td>
<td>3/8 Sc + Ac</td>
</tr>
</tbody>
</table>

two variables (see sections 4.4 and 4.2 for the observations and sunshine measurements). The considered period is May 1995 - April 1997 with synoptic observations at 9, 12, 15 and 18 UTC (2924 observations). The sunshine duration was integrated from 20 minutes before to 20 minutes after the observation time.

Even maximum sunshine duration (10 minutes out of 10) corresponds still to various cloud coverages (635 observations, Tab. 8.2). Low clouds are often present in the sky, but no important fraction of middle altitudes clouds.

Table 8.2: Synoptic clouds observations at Payerne for maximum sunshine duration (10/10 minutes, 20 minutes before to 20 minutes after the observation). Sc: stratocumulus, Cu: cumulus, Ac: altocumulus, Ci: cirrus.

<table>
<thead>
<tr>
<th></th>
<th>Low clouds</th>
<th>Middle clouds</th>
<th>17%</th>
</tr>
</thead>
<tbody>
<tr>
<td>Not ext. Cu</td>
<td>41%</td>
<td>Various Ac</td>
<td>15%</td>
</tr>
<tr>
<td>Sc</td>
<td>17%</td>
<td>Covering Ci</td>
<td>2 cases</td>
</tr>
</tbody>
</table>

Zero sunshine duration (0 minutes out of 10) are observed in 753 cases. The cloud coverage was 7/8 or more in 82% of the observations. Stratocumulus and stratus clouds covered the sky in most cases. Opaque altocumulus were also present. Since we selected data without rain, nimbostratus are not present.

The relation between sunshine duration and cirrus is not straightforward. Observations with 7/8 cirrostratus coverage may lead to maximum (May 24, 1995 at 9 UTC) as well as minimum of sunshine (March 21, 1996 at 9 UTC). It is not very surprising that cirrostratus clouds may have various influences on the sunshine. However, this difference is perhaps partly explained by the sunshine measurement method (threshold of 200 W/m²).

As a conclusion, we note that sunshine duration measurements can be only used with confidence to determine overcast sky data. The clear-sky data have to be extracted in combination with a complementary test, such as based on the ratio direct/global (see chapter 5).
8.3 Overcast sky and UV-Biometer data

The effect of an overcast sky on erythemal UV radiation was estimated with data at Davos between May 1995 and December 1996. Overcast data were extracted by using sunshine duration from the ANETZ. The criterion was zero sunshine for 30 minutes before and after the measurement. Data recorded less than one hour after rain fall stopped were excluded to avoid possible disturbance by water on the dome. We consider ten minute averages data.

The data were normalized to 300 DU and 1 AU with the normalization procedure described in section 5.4 (see plot of the non-normalized data set in Renaud et al., 1997). Results are presented separately for days with and without snow at Davos and in the neighbourhood (see chapters 5 and 7 for the snow conditions). We assume no systematic difference between aerosol optical properties in the clear-sky and overcast sky cases. Variations in the aerosol properties are in any case negligible compared to the cloud effect.

![Figure 8.2: Normalized global UV-Biometer data on days with a snow free surface (left) and with a snow covered surface (right) as a function of the zenith angle. Clear (upper cluster and solid curve) and overcast sky (lower cluster and dashed curve). May 1995 - December 1996 at Davos. Parameterization of the curves with formula: $\log(UV) = a_0 + a_1 m + a_2 m^2 + a_3 m^3 + a_4 m^4 + a_5 m^5$ where $m = 1/\cos\theta$ and $\theta$ is the solar zenith angle in degrees. Clear-sky data without snow: $a_0 = 2.863$, $a_1 = -6.004$, $a_2 = 2.706$, $a_3 = -0.800$, $a_4 = 0.128$, $a_5 = -0.008$. Overcast data without snow: $a_0 = -0.167$, $a_1 = -2.579$, $a_2 = 0.401$, $a_3 = -0.022$, $a_4 = 0$, $a_5 = 0$. Clear-sky data with snow: $a_0 = 2.274$, $a_1 = -4.519$, $a_2 = 1.540$, $a_3 = -0.335$, $a_4 = 0.038$, $a_5 = -0.002$. Overcast data with snow: $a_0 = -0.771$, $a_1 = -1.503$, $a_2 = 0.114$, $a_3 = 0$, $a_4 = 0$, $a_5 = 0$.

An overcast sky on days with a snow free surface at Davos leads to a reduction of the UV-Biometer data to values which averages $30 \pm 1\%$ of the clear-sky value at $30^\circ$ solar zenith angle, $37 \pm 1\%$ at $40^\circ$ and $41 \pm 1\%$ at $70^\circ$ (Fig. 8.2, left). The transmission is larger on days with a snow covered
surface, due to multiple reflections between the surface and the cloud cover. At Davos and on days with snow, the irradiance is reduced in average to 40 ± 1% of the clear-sky value at 40° and 61 ± 1% at 70° (Fig. 8.2, right).

On days with a snow covered surface as well as on days without snow, the impact of overcast cloud layer is highly variable due to the variability of the optical properties and geometry of the clouds. The mean transmission is found to increase significantly with increasing solar zenith angle (similar to Schafer et al., 1996). This is due to the fact that global radiation has a stronger zenith angle dependence on clear-sky days than on overcast days, the ratio overcast sky/clear-sky is then increasing with the zenith angle. The transmission of diffuse radiation would, however, decrease with increasing zenith angle (see e.g. results of Bener, 1964).

Almost no direct radiation is transmitted through clouds and the diffuse radiation is only slightly attenuated (Fig. 8.3). On overcast days, direct and diffuse irradiance are redistributed into diffuse downward and diffuse upward irradiance. The large cloud optical depth leads to the attenuation of direct radiation down to almost disappearance. Note that the ratio direct/diffuse is almost constant on overcast days (around 0.005, see difference between log(direct) and log(diffuse) on Fig. 8.3, right).

![Figure 8.3: Normalized horizontal projected direct (lower cluster) and diffuse UV-Biometer data (upper cluster) as a function of the zenith angle. Clear-sky (left) and overcast sky (right). May 1995 - December 1996 at Davos on days with a snow free surface.](image-url)

A relation between cloud transmission and cloud thickness (determined from radiosondes and observation of the cloud base altitude) has been estimated for stratus clouds at Payerne. An increase of 100 m in cloud thickness led to a decrease of about 6.5% in erythemal UV-Biometer irradiance (44 data points with cloud thickness between 100 and 1000 m; for further information see Röösli, 1997).
8.4 Radiative transfer in clouds

Comparisons between radiative transfer calculations and measurements give further information about the effect of a cloud layer on erythemal UV radiation at Davos. Sensitivity calculations are first performed to estimate the impact of varying altitude and geometrical thickness of a reference cloud layer.

8.4.1 Modelling of the cloud optical properties

Various parameters are used in radiative transfer models to describe optical properties of clouds. Except for Monte-Carlo simulations, models have to assume horizontally homogeneous plane-parallel cloud layers.

In the radiative transfer model TUV, the optical properties of a cloud layer are defined by the altitudes of the bottom and top of the layer, the asymmetry factor \( g_d \), the single scattering albedo \( \omega_{0,cl} \) and the optical depth \( \tau_{cl} \) (independent on the wavelength). One can also determine the optical parameters from liquid water content and droplet radius distribution.

Spinhirne and Green (1978) used \( g_d = 0.86 \) and \( \omega_{0,cl} = 1 \) in the UV range. Hu and Stamnes (1993) gave a parameterization of the asymmetry factor, single scattering albedo and extinction parameter as a function of the wavelength and of the equivalent radius of the droplet. The range of \( g_d \) was 0.85 - 0.9 and \( \omega_{0,cl} \) was larger than 0.999. Standard values for the UV range used in this work are \( g_d = 0.85 \) and \( \omega_{0,cl} = 0.999 \). In the UV range, scattering in the cloud layer is the most important effect (almost no absorption).

Sensitivity to the cloud altitude and thickness was studied with TUV by comparing the effect of a cloud layer with optical depth \( \tau_{cl} = 50 \) between 3 and 4 km a.s.l., between 6 and 7 km a.s.l. and between 3 and 6 km a.s.l. (total ozone amount 300 DU, aerosol optical depth \( \tau_a(340 \text{ nm}) = 0.2 \), altitude of Davos). The calculated global irradiance at ground level increases with increasing cloud layer altitude (maximum 2%) and decreases with increasing thickness (maximum -2.5%). Variations in the cloud altitude and thickness lead to variations in the mean photons optical path length. These differences are small relative to other uncertainties in data and calculations. In the following, we assume a cloud layer between 3 and 4 km a.s.l.

To compare measurements with radiative transfer calculations, ozone and aerosol optical properties have to be known. Since the aerosol properties cannot be measured on cloudy days, we will assume for Davos typical mean properties for summer \( (g = 0.7, \omega_0 = 0.99, \tau_a(340 \text{ nm}) = 0.2) \) and winter \( (g = 0.7, \omega_0 = 0.99, \tau_a(340 \text{ nm}) = 0.05) \) (see chapter 6). The surface albedo is assumed to be 0.03 for days with a snow free surface and 0.5 for days with snow (see chapter 7). Ozone measurements at Arosa are mostly not available for overcast days. In general, we use data on days with at least one
total ozone measurement when using the normalization procedure. Comparisons on specific days were made after estimation of a reasonable total ozone amount in case of missing measurement.

Note that radiative transfer in a cloud cover is extremely idealized with a model such as TUV. The cloud layer optical depth determined by comparison with measurements represents an "equivalent homogeneous plane-parallel cloud optical depth" corresponding to the attenuation observed in the measurements. It cannot be considered as the real optical depth of the cloud.

### 8.4.2 Cloud layer optical depth at Davos

The optical depth $\tau_{cl}$ of a cloud layer has a large impact on the global erythemal UV irradiance reaching the ground (Fig. 8.4).

![Figure 8.4](image)

**Figure 8.4:** Calculated global erythemal UV radiation $UV_{CIE}$ at Davos as a function of the solar zenith angle and cloud layer optical depth $\tau_{cl}$. Summer aerosol properties at Davos (see text) and 300 DU. TUV calculations with DISORT method.

Equivalent cloud layer optical depth $\tau_{cl}$ between 8 and 120 embraces the normalized global UV-Biometer data (300 DU, 1 AU) at Davos with overcast sky and a snow free surface (Fig. 8.5, summer aerosol properties). These extremes correspond to a transmission of global irradiance which ranges between $64\% - 70\%$ ($\tau_{cl} = 8$) and $8\% - 10\%$ ($\tau_{cl} = 120$). The transmission decreases slightly with increasing solar zenith angle, to reach a minimum around $65^\circ$ and increases again for larger zenith angles. This zenith angle dependence is much smaller and has another pattern than the dependence observed in the data. Thus, the transfer in a plane-parallel homogeneous cloud layer and in
8.4. RADIATIVE TRANSFER IN CLOUDS

A real cloud layer seems to disagree. However, the estimates based on the measurements overcast days may be questionable due to the high variability in the data.

Note that the estimated transmissions agree well with the two-stream approximation given by Bohren (1987) for a solar zenith angle $\theta = 0$ and a nonabsorbing medium. In this approximation the transmissivity $T$ through an uniform cloud layer is related to the scaled optical thickness $r^* = (1 - g_d)\tau_d$ with $T = 2/(2 + r^*)$. In our case, we have $T = 63\%$ for $\tau_d = 8$ and $T = 10\%$ for $\tau_d = 120$, which can be compared with 64 - 70\%, respectively 8 - 10\%. As a consequence, the influence of the atmosphere outside the cloud seems to be small at Davos. This is probably due to the fact that the amount of tropospheric ozone is rather small over Davos (see e.g. Frederick and Lubin, 1988 for the influence of tropospheric ozone).

![Figure 8.5: Normalized global UV-Biometer data without snow at Davos (points). Erythemal irradiance with equivalent cloud optical depth $\tau_d = 8$ (thick curves) and $\tau_d = 120$ (thin curves). TUV calculations with DISORT method and summer aerosol properties, see text. $U_{CIE}$ (solid curves) and 0.5 $U_{CH}$ (dashed curves).]

8.4.3 Cloudiness and high surface albedo

Optically thick clouds have a higher reflectivity than optically thin clouds. In case of a snow covered surface, multiple reflections between ground and the cloud bottom are thus increasing with cloud optical depth.

The average increase of global UV-Biometer radiation at Davos due to snow cover for overcast conditions is 43 ± 5\% at 40° and 87 ± 5\% at 70° (see Fig. 8.2). Because of the high variability, these values are rough estimates of the effect, but are clearly much higher than the increase estimated in clear-
sky conditions (about 17% at 40° and 27% at 70°, see chapter 7). The large zenith angle dependence might be a combination of the influence of the cloud layer (see above) and of a non-lambertian behaviour of the reflection at the ground (see discussion in chapter 7). However, the variability is very large and further interpretations are difficult.

Radiative transfer calculations for days with a snow cover (surface albedo 0.5) and cloud layer with optical depth between 8 and 120 do not fit the normalized global UV-Biometer measurements as well as on days without snow (Fig. 8.6, winter aerosol properties). The model seems to underestimate the data. However, the fit is not too bad, if we consider the various uncertainties, such as value of the surface albedo, interaction with the orography, assumption in calculations and eventual systematic differences between winter and summer clouds properties.

Figure 8.6: Normalized global UV-Biometer data with snow at Davos (points). TUV calculations with DISORT method with cloud optical depth \( \tau_{cl} = 8 \) (thick curves) and \( \tau_{cl} = 120 \) (thin curves). \( UV_{CIE} \) (solid curves) and 0.5 \( UV_{CH} \) (dashed curves). Winter aerosol properties, see text.

The increase of global erythemal irradiance due to a snow covered surface under an overcast condition in winter at Davos (winter aerosol properties and surface albedo = 0.5) is estimated to be 40 - 41% with \( \tau_{cl} = 8 \) and 77 - 80% with \( \tau_{cl} = 120 \). These large values illustrate the very important interaction between high surface albedo and cloud layer, as well as its dependence on the cloud optical depth. Only a small zenith dependence is observed in the calculations. Part of this dependence is probably due to a shifting of the spectrum in the erythemal weighted irradiance combined with a wavelength dependent transmission (see last section of this chapter).
8.5 Variability of the influence of clouds

In this section, we analyse the variability of the irradiance measured under a cloudy sky and its relation to the cloud type.

8.5.1 Variability of an overcast sky

The attenuation due to an overcast cloud cover is highly variable between days as well as during the day (examples on Fig. 8.7). This variability is induced by turbulent air motion and the convection induced by radiative cooling at the cloud top. An equivalent cloud optical depth $\tau_d$ can be estimated for each measurement. As an example, $\tau_d$ ranges between 16 and 93 on August 14, 1995 from 10 to 15 UTC (see Fig. 8.8).

Despite overcast cloud cover and high multiple scattering, the diffuse irradiance has a large temporal variability. Although very small values, variations are also observed in the direct measurements. They are related to the variations in the transmissivity of the cloud in the direction of the sun. As an example, see the larger values before 12:30 and the lower values after 14:30 on August 14, 1995 (Fig. 8.9). Comparison between measurements of the direct and diffuse erythemal radiation give further information about the spatial inhomogeneity of the cloud optical depth. If we measure more radiation in the viewing angle of the instrument (8 degrees for the UV-Biometer at Davos) than diffuse would be in the same solid angle, a spatial inhomogeneity is detected in the cloud layer optical depth (e.g. between 11:00 and
Figure 8.8: Global UV-Biometer measurements (left scale, solid curve) and equivalent cloud layer optical depth $\tau_{cl}$ (right scale, dashed curve) at Davos on August 14, 1995 between 10 and 15 UTC. TUV calculations with DISORT method (300 DU, typical summer aerosol properties).

Figure 8.9: Diffuse (solid curve) and direct (dotted curve) UV-Biometer measurements at Davos on August 14, 1995 between 10 and 15 UTC. Note the logarithmic scale on the y axes.

12:00 on August 14, 1995. If the values are similar, the optical depth in the direction of the sun is similar to the average cloud optical depth (e.g. between 12:30 and 14:00 on August 14, 1995).

8.5.2 Broken clouds and enhancement over clear-sky values

The temporal variability of the measured radiation is very high with a broken sky (see e.g. Fig. 8.1 for July 23, 1995). The larger variations are usually due to presence or absence of clouds between the sun and the observer.

Many studies have been conducted to analyse the effect of inhomogeneities
in the cloud cover on the irradiance. In the real world, the shape as well as the content of clouds are inhomogeneous. Radiative transfer in broken clouds remains a major issue in the theory of atmospheric radiation. The effect of scattered clouds on radiation can be studied with Monte-Carlo simulations. Inhomogeneous clouds are generated by distributing clouds with simple geometrical shapes and different sizes over a limited area or as fractal clouds (see e.g. Borde and Isaka, 1996 for results and references, and Loeb et al., 1997 for the problem related to satellite observations).

The radiation measured on ground may be enhanced by reflection of radiation on cloud sides. The most impressive enhancements lead to a measured value larger than the clear-sky value (see section 3.6).

Enhancements have been observed by many authors, most of them with isolated large cumuli in the sky. An enhancement over clear-sky values is usually characterized by an increase of diffuse radiation without variation of direct radiation (sun not obscured by clouds). In all cases, the effect is observed and quantitatively estimated on very few days. Bener (1964) already noticed such enhancements. Mims and Frederick (1994) observed enhancement up to 30% of the clear-sky UV-B at 310 nm, especially when both cloud and sun were near the zenith. Schafer et al. (1996) recorded enhancements of integrated UV-B radiation (transmission larger than 1 with empirical clear-sky reference) up to 11%, all with a sun not obscured. Harshvardhan (1997) found enhancement of UV-A and UV-B irradiance on 7 days in summer 1995, with significant enhancements durations up to 30 minutes. No published results have been found about enhancements by distributed broken clouds.

Similar enhancements of diffuse irradiance with or without reaching a value larger than the clear-sky value are also observed in the Swiss UV-Biometer data set. It is, however, difficult to detect such occurrences, as they have to be distinguished from an increase of diffuse irradiance by redistribution of direct into diffuse irradiance (sun obscured).

Comparisons between measurements and a statistical clear-sky reference (no-snow and snow models, see chapter 5) allows to study possible enhancements at Davos.

UV-Biometer measured values at Davos between May 1995 and April 1996 are often larger than the clear-sky reference, with relative differences up to 37%. Since it is difficult to separate possible bias in the clear-sky reference value from real enhancements, we consider only relative differences larger than 10%. Such enhancements are related to solar zenith angle smaller than about 50° (or ratio direct/global between 0.3 and 0.5) on days without snow (Fig. 8.10). They are mainly observed during periods of scattered cloud covers (for instance stratocumulus). Enhancements for large zenith angles are then not expected because cloud cover is always 100% at high zenith angles (no more direct).
8.5.3 Equivalent optical depth and broken sky

Modelling of enhancement is not possible with a plane-parallel model as it assumes horizontal homogeneity. However, the optical depth of an equivalent homogeneous plane-parallel cloud layer gives some information about the variability in the effect of the clouds on the radiation. Note that this information can only be partial, as the equivalent cloud optical depth is expected to increase with increasing solar zenith angle (mean cloud cover increase with increasing zenith angle).

As an example of the variability in the equivalent cloud optical properties with various cloud types, we show measurements on July 8, 1995 at Davos and the related equivalent cloud optical depths $\tau_{cl}$ (Fig. 8.11 and 8.12). No clouds disturbed the measurements up to around 9:30 UTC (positive values of $\tau_{cl}$ due to an overestimation of the clear-sky irradiance by TUV). Some cumulus clouds led to an attenuation of direct irradiance between 9:30 and 10 UTC (equivalent cloud layer optical depth $\tau_{cl}$ around 10). We have some enhancements over clear-sky reference between 10 and 11 UTC (may be an underestimation of the clear-sky reference). Just around 12 UTC, irradiance is larger than the clear-sky value (could be an enhancement due to reflection on clouds, see the increase of diffuse radiation). Small perturbations are then observed until the arrival of a larger cloud coverage after 12:30 UTC. This cloud coverage is partly scattered and has a maximum equivalent optical depth $\tau_{cl} = 33$.

The integration time is important in the detection of enhancements. These occurrences are usually short in time and may be filtered out by the integration. A crucial part remains also clearly in the choice of the clear-sky...
8.5. VARIABILITY OF THE INFLUENCE OF CLOUDS

reference.

Figure 8.11: Global (solid thin curve), direct (dotted curve) and diffuse (dashed curve) UV-Biometer measurements at Davos on July 8, 1995 (299 DU). TUV clear sky calculations with DISORT method (solid thick curve).

Figure 8.12: Equivalent cloud layer optical depth at Davos on July 8, 1995. Enhancements over the clear-sky value (points under the 0 line). TUV clear sky calculations with DISORT method.

8.5.4 Variability and cloud type

Each cloud interacts with radiation on its own way. However, the behaviour of the short-term temporal variability in global, direct and diffuse irradiance is typical for the different cloud types. As an example, we observe a large variability in direct and diffuse irradiance with stratocumulus clouds, and strong and short in time variability in direct but not in diffuse irradiance with low cumulus clouds (see Fig 8.13).

A non-parametric "cloud variability estimator" \( cve \) is proposed in Appendix D to relate variability in irradiance and cloud type on the basis of measurements and observation at Payerne. Such a method combined with
a discriminant function analysis (separating groups) may be used to better define the relation between variability and cloud type.

8.6 Wavelength dependence

Clouds are expected to introduce only a weak wavelength dependence in the UV. However, a wavelength dependence is observed in measurements and modelling of irradiance reaching the ground.

8.6.1 Measurements and physical processes

Blumthaler et al. (1994) found a larger transmission of clouds for global erythemal UV than for global shortwave irradiance at Jungfraujoch (3576 m a.s.l.). The ratio between reduction of erythemal and shortwave diffuse irradiance under overcast conditions was around 0.36. The influence of the screening of the direct sun led to reduction to the level 62% for global erythemal and 47% for global shortwave irradiance, which corresponds to doubling the reduction ratio to 0.76. The difference is due to the fact that the contribution of the diffuse irradiance in erythema range to the global is much more important than for shortwave irradiance.

At a low altitude station in a large urban area, Frederick et al. (1993) found that cloudy skies provided essentially the same attenuation of irradiance as measured by a Robertson-Berger instrument and a pyranometer. Measurements at longer UV wavelengths (Eppley ultraviolet radiometer 300-380 nm), implied however less attenuation than in pyranometer data. The
authors suggest that the discrepancy between these results could arise from
different sensitivities between Robertson-Berger and Eppley instruments to
tropospheric ozone.

Frederick and Erlick (1997) noted that the measured cloudy-sky irradiance,
expressed as a fraction of the clear-sky value, decreased with increasing
wavelength from 350 to 600 nm (measurements at Palmer Station, Antarctica
and Ushuaia, Argentina).

Spinhirne and Green (1978) deduced from discrete ordinate radiative trans-
fer calculations that, at wavelengths larger than 300 nm, the ratio of UV
transmission to total shortwave transmission through the atmosphere is insen-
sitive to changes of cloud height, cloud scattering parameters and surface
albedo (for realistic values), but is dependent on cloud thickness. For wave-
lengths around 300 nm the absorption by tropospheric ozone introduces also
a sensitivity to cloud height and surface albedo.

The observed wavelength dependence is due to a coupling between the
strongly wavelength dependent Rayleigh scattering and the wavelength in-
dependent reflection of upward moving radiation from the base of the cloud
back to the ground (Frederick and Erlick, 1997). A cloud whose albedo is in-
dependent of wavelength will still have a wavelength-dependent effect on UV
and shortwave radiation at the ground. Consider the downward radiances
at two wavelengths $\lambda_L$ and $\lambda_S$ where $\lambda_L > \lambda_S$ as illustrated in Fig. 8.14.
The fraction of radiation backscattered under a given altitude is greater at
$\lambda_S$ than at $\lambda_L$ (Rayleigh scattering). Under clear conditions most of the
backscattered upward radiation will escape the atmosphere, which is not
the case below a cloud. The result is that cloudy skies appear to have less
attenuation as wavelength decreases, at least as long as absorption is absent.

The larger the ground albedo, the greater the upwelling irradiance from
the surface becomes relative to the contribution from Rayleigh scattering. If
the ground albedo is wavelength independent, or has a much weaker spec-
tral variation than Rayleigh scattering, the albedo reduces the wavelength
dependence below a cloud.

For wavelengths affected by the ozone, Rayleigh scattering would be still more important, but the presence of ozone below the cloud reduces the enhancement towards shorter wavelengths. Furthermore, ozone in the cloud will absorb the multiply scattered radiation and the albedo and transmission of a cloud will no longer be independent of wavelength.

Seckmeyer et al. (1996) found a wavelength-dependent transmittance of a cloud layer between Zugspitze (2964 m a.s.l.) and Garmish-Partenkirchen (720 m a.s.l.), ranging between 45% in the UV-A and 60% in the UV-B ranges. The physical interpretation of the results obtained with this particular cloud was detailed in Kylling et al. (1996). The real cloud transmission (downward flux below cloud/downward flux at top of cloud) exhibits only a very small wavelength dependence, which is due to the small wavelength dependence of the Mie scattering of radiation within the cloud (dependent on the droplet-size distribution). The cloud will increase the upward flux of photons above the cloud. Some of these photons will be backscattered in the downward direction. Hence, on the cloudy day, there will be an increased downward flux of radiation above the cloud. The overall increase in the magnitude of the downward flux on the cloudy day is more pronounced at the cloud top than at Zugspitze. The wavelength dependence is caused by the dominant Rayleigh scattering above 320 nm, where the flux increases with decreasing wavelength, and the ozone absorption, primarily below 320 nm, where the absorption reduces the increase in the downward flux, ultimately reversing the slope effect with wavelength.

8.6.2 Wavelength dependence at Davos

The wavelength dependence of the influence of clouds can also be estimated for Davos by comparing erythemal UV radiation measurements and shortwave pyranometer measurements from ANETZ.

As expected, the short term temporal variability due to a scattered cloudy sky is higher in the shortwave irradiance than in the erythemal irradiance, since the part of direct irradiance in global shortwave is much larger than in global erythemal (example on Fig. 8.15).

The attenuation of the pyranometer data due to an overcast sky averaged 25 - 27% at Davos (Fig. 8.16). This estimate can be compared with the average reduction to 30 - 40% of the clear-sky UV-Biometer data (see above). This difference between attenuation in the UV and shortwave ranges as shown in Fig. 8.17 can be explained by the mechanism described above. The ratio between transmission in erythemal and shortwave ranges between about 0.8 at 40° and 0.5 at 70°, which is larger than the value estimated by Blumthaler et al. (1994) at Jungfraujoch (about 0.36).
8.6. WAVELENGTH DEPENDENCE

Figure 8.15: Relative difference between successive total shortwave pyranometer (solid curve) and UV-Biometer (dashed curve) measurements (10 minutes averages data) on July 8, 1995 at Davos.

Chapter 8 Summary

The influence of clouds on erythemal irradiance shows a large diversity. UV-Biometer data under overcast conditions are reduced to a level which ranges between 8% and 70% of clear-sky value with average around 30 - 40% on days with a snow free surface at Davos. The equivalent cloud optical depth $\tau_d$ (homogeneous plane-parallel cloud layer between 3 and 4 km a.s.l.) ranges between 8 and 120. The increase of erythemal UV-Biometer irradiance due to multiple reflection between a snow covered surface and an overcast cloud cover ranges between 40% ($\tau_d = 8$) and 80% ($\tau_d = 120$) at Davos.

The erythemal irradiance varies strongly on days with a broken as well as an overcast cloud cover. A corresponding series of equivalent cloud layer optical depths can be determined. Difficulties appear in the extraction of enhancements over clear-sky value, due to the sensitivity to the clear-sky reference. Such enhancements seem, however, to be more likely observed at Davos for the smaller solar zenith angles.

Global shortwave irradiance is more attenuated by an overcast cloud cover than erythemal UV irradiance, and its variability is larger in broken clouds. The wavelength dependence measured at ground level is primarily a function of the interaction of the cloud layer with the surrounding atmosphere and the underlying surface.
Figure 8.16: Global total shortwave pyranometer measurements on days with a snow free surface as a function of the solar zenith angle. Clear (upper cluster, solid curve) and overcast (lower cluster, dashed curve) skies between May 1995 and December 1996 at Davos. Parameterization of the curves (pyranometer data) with formula: \( \log(UV) = a_0 + a_1 m + a_2 m^2 + a_3 m^3 + a_4 m^4 + a_5 m^5 \) where \( m = 1/\cos \theta \) and \( \theta \) is the solar zenith angle in degrees. Clear-sky data: \( a_0 = 9.672, a_1 = -4.939, a_2 = 3.295, a_3 = -1.260, a_4 = 0.243, a_5 = -0.018 \) Overcast data: \( a_0 = 6.791, a_1 = -1.348, a_2 = 0.206, a_3 = -0.012, a_4 = 0, a_5 = 0 \).

Figure 8.17: Ratio normalized global UV-Biometer / total shortwave [%] as a function of the solar zenith angle. Clear (lower cluster, thin points) and overcast (upper values, thick points) skies at Davos, May 1995 - December 1996.
Chapter 9

UV Index forecast

In this chapter, we present the UV Index (erythema UV level) used to forecast the possible danger of excessive sun exposure in the public. The Swiss operational forecast of the UV Index is then detailed and its accuracy tested.

9.1 UV Index and forecast methods

Incidence of skin cancer strongly increased in industrialized countries in the last decades (see *e.g.* in Scientific American, 1996 for general information and in Slaper et al., 1996 for scenarios for the future). This increase cannot be explained by an increase in UV radiation received at the ground, but has been shown to be due to increased exposure to sunlight (vacation in low latitudes, excessive sunbathing, etc.; van der Leun et al., 1993; Stachelin and Renaud, 1998).

Queensland, Australia, started the first education campaigns on the prevention of skin cancer and the hazards of overexposure to UV radiation (Long et al., 1996). In the mid-1980s, the Australian Radiation Laboratory began monitoring UV radiation and broadcasting the day’s UV dosage in “minimum erythema dosage” units (MED) for all the states’ capital cities during the evening news. Also New Zealand initiated public awareness campaigns along with broadcasting “burn times” reports hourly on the radio. In 1992, Environment Canada began issuing their own UV Index, a next-day forecast of UV exposure (Burrows et al., 1994). All three countries have been very successful at getting the message to the public about the dangers of being in the sun too long and possible consequences for the skin, eyes, and immune system over a prolonged period of time (Long et al., 1996). UV Index forecasts are presently operational in many countries (see state of the art in 1994 in WMO, 1994).

The Canadian UV Index has been defined as forty times the erythemal irradiance [W/m²]. This index has been recommended by WMO and is used
as common index for the forecast of the erythemal UV level (ICNIRP, 1995). A UV Index of 12 corresponds to the irradiance on a typical mid-summer day for equatorial regions. The UV Index has maximum value around 13 at Davos and 10 at Payerne. Some open issues are presently left about the choice of the presentation: with or without cloudiness, temporal integration (around midday, maximum of the day, average over 30 minutes or 1 hour, etc.). However, some consensus seems to have been found (WMO, 1998) that the UV Index should be given as an average over at least 30 minutes, for the maximum of the day.

The models used for the UV Index forecasts are of different types. Generally, they proceed as follow: (1) forecast of total ozone amount, (2) usage of a model to calculate the radiation reaching the ground (with assumptions on the atmospheric optical properties, e.g. aerosol content). An adaptation for the altitude and the cloud coverage is made either during a further step or directly within the model. Note that some countries are using a total ozone climatology instead of a forecast (e.g. France) assuming that the cloud coverage effect dominates most case.

Information about UV Index forecast procedures can be found for instance in Burrows et al. (1994) for Canada, in Long et al. (1996) for the USA, Austin et al. (1994) for UK, and Feister et al. (1996) for Germany. They use meteorological fields and statistical models for the prediction of the total ozone amount. Burrows et al. (1994) and Austin et al. (1994) use a statistical model to retrieve UV Indices, and Long et al. (1996) and Feister et al. (1996) a radiative transfer program (model from Frederick and Lubin, 1988, respectively from Feister, 1994). Note that forecasts models are developed, adapted and improved continuously.

The COST (European COoperation in the field of Scientific and Technical research) action 713 ”UVB-forecast” was created with the goal of a standardized and proven method of providing UV-B forecast for the general public. A comparison between models has been organized in this framework with 6 multiple scattering spectral models, 8 fast spectral models, and 4 empirical models1. Calculation of the UV Index for 106 European clear-sky atmospheric summer conditions have been compared (various aerosol contents, total ozone amounts, surface albedos and altitudes). The agreement among the multiple scattering models was within ± 0.5 units of UV Index values for more than 80% of chosen atmospheric parameters. The fast spectral models had very different agreements (± 1 up to 12 units of UV Index). The results of the empirical models agreed reasonably well with the reference models, but only for the atmospheres for which they have been developed (Koepke et al., 1998). An intercomparison between model outputs and measurements is presently under way (first results discussed in Helsinki in July 1998).

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1The models developed for the station Davos (see chapter 5), with altitude adaptation based on the measurements at Davos and Payerne participated in that COST comparison.
9.2 Swiss UV Index forecast

The SMI produced its first operational UV Index forecast (48 hour forecast for midday) during summer 1996 (see Fig. 9.6; person in charge: P. Eckert, SMI Geneva). The UV Index forecast is presented in the frame of the campaigns to raise public awareness about the danger of excessive sun exposure in collaboration with the Swiss Federal Office of Public Health.

In the Swiss procedure, a clear-sky UV Index is first forecasted for Davos by using the statistical no-snow model presented in chapter 5, with calculated (solar zenith angle, correction Sun-Earth distance), respectively forecasted (total ozone amount, see below) input parameters\(^2\). By applying altitude and cloudiness correction factors (see below), the UV Index is, in a second step, forecasted for 14 regions\(^3\) in Switzerland. Values at 1000 m, 2000 m, and 3000 m are also given for the mountainous regions (see Fig. 9.6).

9.2.1 Total ozone forecast

On time scales of a few days or less the dynamics is the most important factor affecting stratospheric ozone levels (see section 3.3). Total ozone has been shown to be correlated with lower stratospheric temperatures, geopotential heights and winds, and with isentropic potential vorticity (near the tropopause or total column). The total ozone amount \(X_t\) on day \(t\) or the change between successive days \(X_t - X_{t-1}\) are usually forecasted by using forecasted meteorological fields and a statistical model.

Allard et al. (1993) used the correlation between column integrated potential vorticity and TOVS total ozone data and obtained a root-mean-square error of the order of 15 DU in summer and 20 to 25 DU in winter. Poulin and Evans (1994) correlated total ozone amounts with the stratospheric temperature at either 30 or 50 hPa, and the height of the 250 hPa surface. The absolute average per cent differences between model and Brewer measurements at Edmonton were all close to 7% with a standard deviation 5 - 6%.

Burrows et al. (1994) explained about 82% of the variance in total ozone (TOMS measurements over Canada) with 6 predictors (main predictor: climatological ozone amount). The root-mean-square error in the test on an independent data set was about 11 DU (corresponds to a difference in the UV Index of 0.25 in early July for a UV Index around 8 at Toronto). Vogel et al. (1995) also used temperature and geopotential at standard pressure levels (50, 100, 500 and 700 hPa; complement in Spänkuch and Schultz, 1997) and

\(^2\)In summer 1996, the statistical model was based on clear-sky UV-Biometer measurements at Arosa. The model was similar to that for Davos.

explained 64 to 83% of the variance in total ozone over Germany depending on the month. The limit of 20 DU maximum deviation was exceeded only for large errors in the predictors.

Feister et al. (1996) had an uncertainty of ozone forecast of 3 - 6% (regressors: ozone amount at previous day, relative vorticities and temperatures at 5 pressure levels). Long et al. (1996) found root-mean-square errors of about 9 to 14 DU over the Northern Hemisphere with a model taking into account the changes in geopotential height at 100 and 500 hPa, and the changes in temperature at 50 hPa (the change in ozone amount between successive days is forecasted).

A total ozone forecast procedure for Arosa has been developed at LAPETH (contribution from A. Gamma and J. Bader, information in Baumann, 1994). The regression model has time lags and the error is modelled by an autoregressive process AR(1). It is expressed by

\[ X_t = S_t + \sum_{k=0}^{2} \beta_k \, t_{100,t-k} + \sum_{k=0}^{2} \gamma_k \, t_{700,t-k} + \sum_{k=0}^{2} \delta_k \, t_{p2p,v,t-k} + e_t \]

\[ e_t = p_t \, e_{t-1} + \nu_t \]

(9.1)

\[ \nu_t \sim N(0,1) \]

with

\[ t_{100,t} = t_{100,t} - st_{100,t} \]
\[ t_{700,t} = t_{700,t} - st_{700,t} \]
\[ t_{p2p,v,t} = t_{p2p,v,t} - st_{p2p,v,t} \]

(9.2)

where \( X_t \) is the total ozone amount at day \( t \), \( S_t \) is the climatological total ozone amount at time \( t \) (interpolation between monthly averages, taking trends into account). Variables \( t_{100,t} \) and \( t_{700,t} \) are temperatures [K] at the 100 hPa respectively 700 hPa levels. Variable \( t_{p2p,v,t} \) is the pressure at the tropopause level, which is defined by a potential vorticity \( PV \) equal to 2 units\(^4\). Variables \( st_{100,t} \), \( st_{700,t} \) and \( st_{p2p,v,t} \) are the corresponding climatological values. The parameters \( \beta_k, \gamma_k, \delta_k, k = 0, 1, 2 \) and \( p_t \) were estimated with analysis data from the ECMWF (temperatures and pressure at tropopause level) and total ozone values (D101(AD) daily mean measurements) at Arosa (Tab. 9.1).

The operational usage of the model started at the end of 1994 at LAPETH to produce a daily total ozone forecast for Arosa and ended 1997.

\(^4\) The potential vorticity \( PV \) is expressed by (isentropic coordinate form; Holton, 1992)

\[ PV = -g \frac{\partial \theta}{\partial \rho} (\zeta \phi + f) \]

where \( g \) is the gravity, \( \theta \) is the potential temperature, \( \rho \) is the pressure, \( \zeta \phi \) is the vertical component of relative vorticity evaluated on an isentropic surface and \( f = 2\Omega \sin \phi \) is the Coriolis parameter (\( \Omega \): angular velocity, \( \phi \) = latitude). The PV is conserved following the motion in adiabatic frictionless flow. 1 PV unit = 10\(^{-6}\) Km\(^2\) kg\(^{-1}\) s\(^{-1}\).
Table 9.1: Parameters of the total ozone model at Arosa.

<table>
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<th>Value</th>
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</thead>
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<tr>
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</tr>
</tbody>
</table>

During 1995, the absolute error in total ozone was smaller than 56.9 DU (16%), with mean absolute value around 12 DU (3.9%). The residuals had a standard deviation of 4.6%. The main cause for uncertainties has been found to come from the lack of accuracy in the forecasted temperature at the 100 hPa level and from the total ozone climatology (Kramer, 1996).

The SMI is using the model described in Eq. 9.1 and 9.2 in the operational UV Index forecast, but without taking into account the regressive process of the error (practical reason: the measured value $X_{t-1}$ is not available in the +48 hours forecast). In summer 1997, we had 84 days with reliably forecasted values (Fig. 9.1, May 15 - September 5). The relative difference between forecasted value and measurements ranged between -10% and +9%.

![Figure 9.1: Total ozone measurements (D101(AD), solid line and "·") and forecasted values (dashed line and "0") between May 15 and September 5, 1997 at Arosa.](attachment:figure91.png)

The forecast procedure has a good ability to predict the day-to-day changes in total ozone (77% of the changes are well forecasted, see Tab. 9.2). However, the amplitude of the changes is generally smaller than the observed one. This smoothing may be explained by the ECMWF meteorological fields smoother than in the real atmosphere, and the presence of time lags in Eq. 9.1.
Table 9.2: Observed and forecasted day-to-day changes in total ozone amount between May 15 and September 5, 1997 at Arosa. "+" is an increase since the previous day, "-" is a decrease since the previous day.

<table>
<thead>
<tr>
<th></th>
<th>Forecasted</th>
<th>sum</th>
</tr>
</thead>
<tbody>
<tr>
<td>Observed</td>
<td>+</td>
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<tr>
<td></td>
<td>13</td>
<td>3</td>
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<td>-</td>
<td>5</td>
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<tr>
<td></td>
<td>14</td>
<td>19</td>
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<tr>
<td>sum</td>
<td>18</td>
<td>17</td>
</tr>
</tbody>
</table>

In the Swiss UV Index forecast, we assume that the total ozone amount above the altitude of Arosa is constant over Switzerland. This assumption is reliable for most summer days. It would, however, be questionable in spring, where strong and rapid variations are observed over Switzerland. An ozone analysis and forecasting from the ECMWF should be available by the end of 1998 (P. Eckert, personal communication February 1998), and could then be used in the UV Index forecast procedure instead of Eq. 9.1 and Eq. 9.2. Note, however, that an ECMWF ozone forecasted value should probably be corrected to be directly comparable with measurements at Arosa before application in the no-snow model for the UV Index forecast.

### 9.2.2 Altitude attenuation

The difference in erythemal radiation between Payerne and Davos has been shown to vary strongly over the year and between days (section 6.5). In the operational UV Index forecast, we use two linear corrections (boundary layer below the altitude of Davos and "free troposphere" above the altitude of Davos). The clear-sky value at altitude $h$ [m] is obtained as the product of the clear-sky value at Davos by the altitude attenuation factor $AAF(h)$ expressed by

$$AAF(h) = \begin{cases} 
1 - 0.18 \frac{1610 - h}{1000} & \text{if } h \leq 1610 \text{ m} \\
1 - 0.10 \frac{1610 - h}{1000} & \text{if } h > 1610 \text{ m}
\end{cases} \quad (9.3)$$

This factor corresponds to average summer values based on measurements in Payerne and Davos and radiative transfer calculations (Fig. 9.2).

### 9.2.3 Clouds attenuation

UV forecast on cloudy days is complex, as the forecast of the cloudiness is difficult and a cloud attenuation factor $CAF$ can only be a rough approximation for the real impact. Furthermore, the cloud attenuation factor has to
Simple attenuation factors are usually employed in forecast procedures. As an example, Burrows et al. (1994) chose 0.4 or 0.7 in overcast conditions depending on whether precipitation is also forecast. Green et al. (1974) gave the following approximation of the attenuation in global UV-B

\[ CAF(f) = 1 - 0.56 f \] (9.4)

where \( f \) is the fraction of the sky covered by clouds. Other parameterizations are also given in Frederick and Steele (1995) for application to erythemally weighted UV radiation. For instance

\[ CAF(f) = 1.06 - 0.51 f \] (9.5)

for the UV range (Eppley ultraviolet radiometer, 300 - 380 nm) at 12:00 (39% of variance explained). Schafter et al. (1996) gave a parameterization dependent on \( C_{(10)} \) the cloud amount in tenths (UV-B range)

\[ CAF(C_{(10)}) = 1 - 0.1312 C_{(10)} + 0.0287 C_{(10)}^2 - 0.0022 C_{(10)}^3 \] (9.6)

Bais et al. (1993) found that for categories 0/8 to 2/8 the UV irradiance may be considered constant. The following polynomial

\[ CAF(C_{(8)}) = 1 + 0.06 C_{(8)} - 0.02 C_{(8)}^2 \] (9.7)

adequately fit their experimental data (spectral measurements at Thessaloniki, Greece, 50° solar zenith angle) with cloudiness higher than 2/8 (\( C_{(8)} \))
The U.S. National Weather Service (Long et al., 1996) used the formula
\[ CAF = 0.316 + 0.676(\%\text{chance clear}) + 0.58(\%\text{chance scattered}) + 0.41(\%\text{chance broken}) \] (9.8)
where clear = 0-1 tenths, scattered = 2-5 tenths and broken = 6-8 tenths.
The factor \( CAF \) equals 0.316 with an overcast sky.

Most of the operational cloud forecast systems use output of weather forecast models combined or not with statistical models (see e.g. Long et al., 1996). In the operational UV Index forecast in Germany, nowcasting of cloud cover by the forecasters in the morning hours allows reducing uncertainties (Feister et al., 1996). Different approaches were tested in Germany to determine the cloud cover from the total cloud cover or from combinations of the cloud covers at different heights. The least uncertainties were obtained by the sum of clouds at low and medium heights.

In the Swiss operational UV Index forecast, the cloud coverage for the 14 regions is forecasted by forecasters from the SMI. Half of the coverage in high clouds is added to the coverage of low and middle clouds. The following formula is used to adapt the clear-sky values to the cloudy sky final values (P. Eckert, personal communication)
\[ CAF(C_{(S)}) = 1 - 0.0133 C_{(S)}^2 \] (9.9)
where \( C_{(S)} \) is the cloud coverage in eighths (low value for 8/8 coverage since forecasted only in case of a very thick overcast cloud cover; see Fig. 9.3).

### 9.2.4 Accuracy of the UV Index

As soon as clouds are taken into account, a UV Index forecast is often not very reliable. As an example, Feister et al. (1996) noted that the model calculations overestimated the measured radiation by about 10 to 15\% for cloud cover less than 4/8 to 5/8. In the USA, the UV Index was overestimated for lower cloud attenuation factors and underestimated at higher cloud attenuation factors in 1993 (Long et al., 1996). In summer 1994, the UV Index in the USA was within 1 unit for 76\% of the studied cases and within 2 units for 91\%. Forecasts tended to overestimate the UV radiation during cloudy days and underestimate it during clear days.

The Swiss UV Index forecast from summer 1997 was tested by comparison with measurements at Davos and Payerne. The mean measured values around midday (30-minutes average in order to smooth possible peaks) were compared with the UV Index forecasted value. Further information about the time integration and daily plots in Seneviratne (1998).

At Davos (cloudiness of region 11) and Payerne (cloudiness of region 3), 24\% and 19\%, respectively, of the forecasted values agreed with the measured 30-minutes averages (54 days, rounded values, see Tab 9.3 and 9.4).
9.2. SWISS UV INDEX FORECAST

Figure 9.3: Cloudiness adaptation factors $CAF$ as a function of the fraction of the sky covered by clouds. Green et al. (1974) (solid thin curve), Frederick and Steele (1995) (dotted curve), Shafter et al. (1996) (dashed curve), Bais et al. (1993) (dotted-dashed curve) and in the SMI operational UV index forecast in summer 1997 and 1998 (solid thick curve).

This proportion reaches 57%, and 59% when considering the UV Index ± 1. Various potential errors could explain these only partly satisfactory results: uncertainties in the clear-sky model (including errors in the ozone forecast and possible calibration shifts), and inadequate altitude and/or cloud attenuation factors.

At Davos, no day with forecasted or observed 0/8 cloud coverage around midday are observed in summer 1997. Only 5 days had very small coverage which did not influence significantly the irradiance measurements (rather smooth curves, with cloud coverage observations between 1/8 and 4/8 and forecast between 1/8 and 3/8). We note that the clear-sky forecasted values overestimated the measurements on these days by 2 to 8%. This overestimation may be due to aerosol turbidities larger than average or to a bias in the measurements (see the overestimation for the control data set in chapter 5). However, UV Index forecasts remained within 1 UV Index from the measurements (with and without taking into account the clouds).

Clear-sky forecasts and measurements at Payerne on days with a smooth irradiance curve were also within 1 UV Index (8 days with cloud coverage observations between 1/8 and 3/8, and forecast between 0/8 and 4/8). However, a systematic underestimation of the measurements is observed (-2 to -10%). The altitude attenuation factor $AAF(h)$ seems to be too large on these days.

The ratio between 30-minutes averages measurements around midday at Payerne and Davos ranges between 0.9 and 0.99 on the simultaneous clear-
Table 9.3: Contingency table of observed and forecasted UV Index at Davos between May 15 and September 5, 1997.

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<td>11</td>
<td>8</td>
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<td>54</td>
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</tbody>
</table>

sky days (July 16 and 30, and August 10). The underestimation of the clear-sky values at Payerne could then be explained by an unusually low turbidity in the layer between both stations (see section 6.5 for comparison with other values), or by a calibration problem, such as measurements too large at Payerne or too small at Davos.

The forecast error in the UV Index is clearly dependent on the error in the cloud coverage forecast (Fig. 9.4, cloud coverage from SYNOP observations at Davos and Payerne at 12 UTC). The forecasted clear-sky values are sometimes exceeded with scattered cloud coverage, probably due to especially low aerosol turbidities or other errors (altitude factor), rather than real enhancements from reflection on cloud sides. The forecasted cloud cover is generally smaller than the observed cloud cover, especially at Davos, for coverage larger than 5/8. For small coverage, the cloud cover is rather slightly overestimated.

The forecasted UV Index with the observed cloud coverage instead of the forecasted cloud coverage is systematically lower than the 30-minute measured averages (bias of 1.75 UV Index at Davos and 1.16 UV Index at Payerne). Although only partial information on the cloud impact is found in the cloud coverage observations (e.g. no differentiation between clouds at the horizon or between the sun and the station, time factor), the variability in the error is clearly smaller than with forecasted cloudiness.

Considering that forecasted cloud coverage should be taken as a basis for the cloud attenuation factor (with its bias in comparison with the observa-
Table 9.4: Contingency table of observed and forecasted UV Index at Payerne between May 15 and September 5, 1997.

<table>
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<td>54</td>
</tr>
</tbody>
</table>

The cloud attenuation factor seems to be an adequate rough estimation for Davos (Fig. 9.5, left). At Payerne, data with low coverage are clearly underestimated by the clear-sky forecast. This confirms the remark above about the overestimation of the altitude effect between Davos and Payerne in summer 1997.

An adaptation of the cloud coverage factor and altitude effect between Davos and the Plateau have to be considered as soon as additional data are available for validation (much more data required for more reliable algorithm).

Chapter 9 Summary

The UV Index (40 times erythemal irradiance in W/m²) is used as unit in forecasts of the erythemal UV level in campaigns on the prevention of skin cancer and hazards of overexposure to UV radiation (skin cancer, eye disorder, immune system).

In Switzerland, a UV Index forecast is operational during summer since 1996. The forecast algorithm is based on a total ozone forecast (statistical model with ECMWF forecasted meteorological fields as inputs), the clear-sky no-snow model for Davos, an estimated altitude dependence, and an estimated attenuation due to cloudiness. Uncertainties remain in the forecast of the ozone, aerosol conditions and cloud coverage. The UV Index forecast of summer 1997 is only partly satisfactory. Since most of the error is due
to uncertainties in the cloud coverage, it should be worthwhile to reconsider including the expected clear-sky values as a function of the altitude in the forecast to serve as maximum values or as worst-case for exposure to sunlight (done in 1996, but not in 1997 and 1998).
Figure 9.6: Swiss UV Index forecast for June 5, 1998. Facsimile product from the SMI.
Chapter 10

Conclusion

The purpose of the present work is to improve our understanding of the interaction of UV radiation with real atmospheres. The method is a combined analysis of measured data and radiative transfer calculations in order to estimate the qualitative and quantitative influences of changes in atmospheric ozone content, aerosol particle condition, surface reflectance, altitude and cloud cover.

The reliable UV-Biometer measurements of direct, diffuse and global erythemal UV radiation at Davos and Payerne have been shown to contain extensive information about the absorption and scattering processes in the atmosphere. Radiative transfer calculations helped in the interpretation of the results based on measurements by allowing more detailed studies of individual parameters.

The main results of this research work may be summarized as follows (see also the summaries at the end of chapters 5 to 9):

- The two most influencing parameters on clear-sky days are solar zenith angle $\theta$ and total ozone amount. A statistical model is used to quantify the influence of their variations on erythemal radiation at Davos. The global erythemal irradiance is at $\theta = 30^\circ$ 11.5 times larger than at $\theta = 70^\circ$ with a total ozone amount equal to 300 DU. For comparison, a variation in total ozone from 240 to 360 DU leads to an increase by a factor of 1.4 at $\theta = 30^\circ$. The relationship between decrease in total ozone and increase in erythemal UV radiation strongly depends on the solar zenith angle and on the total ozone amount (see chapter 5).

- The impact of variations in total ozone amount and Sun-Earth distance can be filtered out by a normalization procedure. This allows the study of other influences such as aerosol, surface albedo and clouds (see section 5.4).

- Variation in the aerosol loading influences the distribution of global irradiance into direct and diffuse irradiance. Global erythemal irradi-
ance, however, varies by maximal 5% between days with very low and very high aerosol turbidity at Davos (see chapter 6).

- The atmospheric aerosol content has a large variability in the polluted boundary layer. As a consequence, the relative difference between global erythemal radiation at Davos and at Payerne is also strongly varying (average values around 20 to 40% in summer, see section 6.5).

- The measured erythemal radiation also depends on the ground reflectivity. If the ground is covered by snow on clear-sky days at Davos, erythemal irradiance increases by 15 to 25% due to multiple reflections between the surface and the atmosphere (see chapter 7). This relative increase may reach 80% on overcast days depending on the cloud optical depth (see chapter 8).

- Erythemal radiation varies the most under cloudy conditions. On overcast days with a snow free surface, erythemal irradiance reduces to a level which ranges between 8% (very thick cloud cover) and 70% (thin cloud cover) at Davos relative to clear sky. Under a broken cloud cover, the irradiance varies strongly and may be larger than the clear-sky value because of reflections on cloud edges and sides (see chapter 8).

- The Swiss UV Index forecast procedure operational at the Swiss Meteorological Institute aims at raising public awareness about the risk of excessive exposure to UV radiation. It showed good correspondence on clear-sky days. The most severe limitations are the unreliable forecast of cloud, aerosol loading condition and ozone amount (see chapter 9).

One of the strengths of the present analysis is the existence of reliable direct, diffuse and global UV-Biometer measurements in parallel with independent measurements of atmospheric ozone, atmospheric optical depth and meteorological parameters; all the year round and at a high frequency (2-minute recorded UV-Biometer data). All these data allows an extensive study of scattering and absorption processes. Most of the results of this work are consistent with those obtained in earlier studies, but more specific to the alpine region.

The results based on the analysis of measurements and on model calculations show the strengths of such a combined method. The meticulous control of the instruments allows good confidence in the data. Nevertheless, measurement accuracy, as well as local effects, representativeness, variability and combination of many simultaneous complex physical processes influence the recorded value and the radiative transfer model also has uncertainties. These remarks raise awareness regarding the difficulty to generalize results and findings. Despite inevitable limitations in both measurements and model calculations, however, the estimates can be considered representative for a moderately elevated station in the Swiss Alps (Davos) and for the Swiss Plateau (Payerne) respectively.
The results obtained in this PhD thesis are exploratory and can be used for further analysis.

1. The estimates are based on measurements during limited periods (e.g. one year). The analysis of longer periods is now possible and would allow testing and improving the present approximations and models. The direct determination of trends would obviously require much more years of reliable measurements.

2. Radiative transfer calculations with a more accurate description of the atmosphere, as well as of the aerosol particles optical properties and size and spatial distribution would be interesting for further comparisons with measurements.

3. A detailed investigations of the influence of orography on measured irradiance in a complex terrain would be of great interest for any station in the Alps. Monte-Carlo calculations could allow analysing this effect from a simple idealized orography with a snow free surface up to a complex terrain with a snow covered surface.

4. A further analysis of erythemal and shortwave measurements combined with intensive cloud observations would be interesting to better define the attenuation and variability of radiation for different cloud types.

5. The operational UV Index forecast also showed that further investigation of the cloud influence is clearly needed (forecast of the cloud coverage, modelling of the attenuation). A forecast of the aerosol turbidity would also improve the UV Index forecast on the Swiss plateau. Furthermore, the altitude adaptation for high level stations could now be estimated by comparing measurements at Davos and Jungfraujoch. Ameliorations in the ozone forecast would also improve results in clear-sky conditions.

6. The results are based on measurements with a horizontal receiver, but the consideration of an inclination of the receiver is very important when applying results to the information about exposure of people to solar UV radiation. Furthermore, discussions between physicians and physicists could probably come out on relevant results concerning the choice of an adequate UV Index and of its broadcasting.

UV radiation has not yet unveiled all its secrets! It is rewarding going further into this research domain, which is interesting from the viewpoint of atmospheric science and for public concerns regarding the danger of excessive sun exposure.
Appendix A

Astronomical calculations

A.1 Variation of the extraterrestrial irradiance

As a consequence of the variation of the Sun-Earth distance during the year, the intensity of the solar energy arriving at the top of the atmosphere varies by about ±3.5%. The result of the Fourier analysis of Spencer (1971) is used as a WMO standard (WMO, 1985a) for the Sun-Earth distance correction \( d_c \)

\[
d_c = 1.000110 + 0.034221 \cos y + 0.00128 \sin y + 0.000719 \cos 2y + 0.000077 \sin 2y \quad (A.1)
\]

where \( y = 2 \pi (j-1)/365 \) with \( j \) = julian date. Note that \( d_c = (R_j/R_0)^2 \), where \( R_j \) is the distance between the sun and the earth on day \( j \) and \( R_0 \) is the mean Sun-Earth distance. More accurate algorithms have been developed; an example is from Michalsky (1988) as a Fortran program.


Input:
- year (e.g. 1986.0)
- dayno of year (e.g. Feb.1 = 32.0)
- hour UT fractions e.g. 8:30 AM eastern daylight is 8.5 + 5 - 1
- latitude north positive
- longitude east positive

Output:
- radius Sun-Earth distance [AU]
- azimuth east from north, south is 180deg
- elevation includes refraction

SUBROUTINE Michalsky(year,dayno,hour,lat,long, & azimut,elevation,radius)
IMPLICIT NONE

INTEGER, INTENT(in) :: year, dayno
REAL, INTENT(in) :: hour, lat, long
REAL, INTENT(out):: radius, azimut, elevation

INTEGER :: delta, leap
REAL, PARAMETER :: pi=3.1415927, twopi=6.2831853, d2r=0.017453293
REAL :: jd, time, mnlong, mnanom, eclong, oblqec, gmst, lmst
REAL :: az, el, refrac, ra, dec, ha, num, den, latr

! get the current Julian date (actually add 2'400'000 for jd)
delta = year - 1949
leap = INT(delta/4)
jd = 32916.5 + delta*365.0 + FLOAT(leap + dayno) + hour/24.
time = jd - 51545.0 ! 2'451'545.0 = 1 jan 2000 12:00 GMT

! calc ecliptic coordinates
! mean longitude between 0 and 360 degrees
mnlong = 280.460 + .9856474*time
CALL angl36(mnlong)
! mean anomaly between 0 and 360 degrees
mnanom = 357.528 + .9856003*time
CALL angl36(mnanom)
mnanom = mnanom*d2r
! compute ecliptic longitude and obliquity of ecliptic in radians
eclong = mnlong + 1.915*sin(mnanom) + 0.020*sin(2.*mnanom)
CALL angl36(eclong)
oblqec = 23.439 - .0000004*time
eclong = eclong*d2r
oblqec = oblqec*d2r

! calculate right ascension and declination in radians
num = cos(oblqec)*sin(eclong)
den = cos(eclong)
ra = atan(num/den)
IF (den.LT.0) THEN
  ra = ra + pi
ELSEIF (num.LT.0) THEN
  ra = ra + twopi
ENDIF
dec = asin(sin(oblqec)*sin(eclong))

! calculate sun-earth distance
radius=1.00014 - 0.01671*cos(mnanom) - 0.00014*cos(2.*mnanom)

! calculate Greenwich mean sidereal time in hours
! hour not changed to sidereal since 'time' includes fractional day.
gmst = 6.697375 + .0657098242*time + hour
gmst = MOD(gmst,24.)
IF (gmst.LT.0.) gmst = gmst + 24.
! calculate local mean sidereal time in radians
lmst = gmst + long/15.0; lmst = MOD(lmst,24.)
IF(lmst.LT.0.) lmst = lmst + 24.
lmst = lmst*15.*d2r

! calculate hour angle in radians between -pi and pi
ha = lmst - ra
IF (ha.LT.-pi)ha = ha + two pi
IF (ha.GT.pi) ha = ha - two pi

! calculate azimuth and elevation
latr = lat*d2r ! ** modifies argument, so use dummy
el = asin(sin(dec)*sin(latr) + cos(dec)*cos(latr)*cos(ha))
az = asin(-cos(dec)*sin(ha)/cos(el))
IF(sin(dec) - sin(el)*sin(latr).GE.0.) THEN
  IF (sin(az).LT.O.) az = az + two pi
ELSE
  az = pi -az
ENDIF
! calculate refraction for US std.atm.
! note that 3.51561 = 1013.2mb/288.2K
el = el/d2r
IF (el.GT.-.56) THEN
  refrac=3.51561*(.1594+.0196*el+.00002*el**2)/ &
  (1. + .505*el + .0845*el**2)
ELSE
  refrac = .56
ENDIF
elevation = el + refrac ! elevation in degrees
azimuth = az/d2r
RETURN

CONTAINS

SUBROUTINE angl36(w)
REAL, INTENT(inout) :: w
w = MOD(w,360.)
IF(w.LT.0.)w=w + 360.0
END SUBROUTINE angl36

END SUBROUTINE Michalsky

SUBROUTINE angl36(w)
REAL, INTENT(inout) :: w
w = MOD(w,360.)
IF(w.LT.0.)w=w + 360.0
END SUBROUTINE angl36

END SUBROUTINE Michalsky
A.2 Solar zenith angle and local true time

In the present report, the solar zenith angle and equation of time (equals difference between mean and true time of the meridian transit) are calculated with the operational algorithm used at the LKO at Arosa (Schill, 1994).

Constants

\[
\begin{align*}
\text{anom}_0 &= 356.241 \text{ (anomaly-constant for the year 1980)} \\
\text{sl}_0 &= 278.828 \text{ (sunlength-constant for the year 1980)} \\
a_1 &= -105.0 \\
a_2 &= 596.2 \\
a_3 &= 4.3 \\
a_4 &= -12.7 \\
b_1 &= -429.3 \\
b_2 &= -2.1 \\
b_3 &= 19.3 \\
\text{clam}_1 &= 0.033423 \\
\text{clam}_2 &= 0.000349 \\
\text{con}_1 &= 0.26179939 \\
\text{con}_2 &= 0.985600267 \\
\text{cond} &= 0.39781 \\
\text{conl}_1 &= 0.985647 \\
\text{conl}_2 &= 0.98567335 \\
\text{cont} &= 0.9856 \\
\text{ufac} &= 0.017453293 \text{ (deg-rad conversion factor)} \\
\text{cycle} &= 1461 \text{ (number of days in a four year cycle)}
\end{align*}
\]

Calculation of the sunlength, anomaly, and declination

Calculation of the ecliptic longitude of the sun (sunlength), the correction for varying velocity of the earth on its orbit (anomaly), and the angle between the ecliptic and the equatorial plan (declination, dec). The input parameters are: year (yyyy), daynumber [1...366] and gmt (UTC time in hour).

\[
\begin{align*}
dy &= \text{year}-1980 \\
\text{leapcy} &= \text{dy DIV 4} \\
\text{addy} &= \text{dy MOD 4} \\
\text{ndleapcy} &= \text{leapcy*cycle}
\end{align*}
\]

\[
\begin{align*}
\text{addy} & \quad \text{ndaddy} \\
-3 & \quad -1095 \\
-2 & \quad -730 \\
-1 & \quad -365 \\
0 & \quad 0 \\
1 & \quad 366 \\
2 & \quad 731 \\
3 & \quad 1096
\end{align*}
\]
A.2. **SOLAR ZENITH ANGLE AND LOCAL TRUE TIME**

\[
\text{ndprev} = \text{ndleapcy} + \text{ndaddy}
\]
\[
\text{dym} = \text{daynumber} + \text{gmt}/24.0
\]
\[
\text{sunlength} = \text{si0} + \text{cons1} \ast (\text{ndprev} + \text{dym})
\]
\[
\text{sunlength} = (\text{sunlength} - \text{trunc} (\text{sunlength}) + (\text{trunc} (\text{sunlength}) \text{ MOD } 360)) \ast \text{ufac}
\]
\[
\text{anomaly} = \text{anom0} + \text{conan} \ast (\text{ndprev} + \text{dym})
\]
\[
\text{anomaly} = (\text{anomaly} - \text{trunc} (\text{anomaly}) + (\text{trunc} (\text{anomaly}) \text{ MOD } 360)) \ast \text{ufac}
\]
\[
\text{ecl} = \text{sunlength} + \text{claml} \ast \sin (\text{anomaly}) + \text{clam2} \ast \sin (\text{anomaly} \ast 2.0)
\]
\[
\text{dec} = \text{asin} (\text{cond} \ast \sin (\text{ecl}))
\]

**Calculation of the equation of time and local true time**

The calculation of the local true time \(\text{woz}\) (in hours) requires following input parameters: year \([\text{yyyy}]\), daynumber \([1..366]\), longitude (\text{long}), and gmt (UTC time in hours).

\[
\text{sl2} = \text{sunlength} \ast 2.0
\]
\[
\text{sl3} = \text{sunlength} \ast 3.0
\]
\[
\text{sl4} = \text{sunlength} \ast 4.0
\]
\[
\text{zgs} = a1 \ast \sin (\text{sunlength}) + a2 \ast \sin (\text{sl2}) + a3 \ast \sin (\text{sl3}) + a4 \ast \sin (\text{sl4})
\]
\[
\text{zgc} = b1 \ast \cos (\text{sunlength}) + b2 \ast \cos (\text{sl2}) + b3 \ast \cos (\text{sl3})
\]
\[
\text{zg} = \text{zgs} + \text{zgc}
\]
\[
\text{zz} = 4 \ast \text{long}/60
\]
\[
\text{wozkorr} = \text{zz} + \text{zg}/3600
\]
\[
\text{woz} = \text{gmt} + \text{wozkorr}
\]

**Calculation of the zenith angle**

The zenith angle \(\text{zen}\) [degree] is calculated as a function of the local true time \(\text{woz}\), the latitude \(\text{lat}\), and the declination \(\text{dec}\).

\[
\text{hrangle} = (\text{woz} \ast 15.0 - 180) \ast \text{ufac}
\]
\[
\text{sunhei} = \cos (\text{lat} \ast \text{ufac}) \ast \cos (\text{dec}) \ast \cos (\text{hrangle}) + \sin (\text{lat} \ast \text{ufac}) \ast \sin (\text{dec})
\]
\[
\text{sunhei} = \text{asin} (\text{sunhei}) / \text{ufac}
\]
\[
\text{zen} = 90 - \text{sunhei}
\]
Appendix B

Optical tables approximations

B.1 Relative optical air mass

Kasten (1966) gave a formula for the air mass $m_r$ at sea level as a function of the zenith angle $\theta$ (in degree) at the observer (astronomical corrected for refraction)

$$m_r = \frac{\cos \theta + 0.15(90 - \theta + 3.885) - 1.253}{1}$$

(B.1)

and a revision of this approximation is presented in Kasten and Young (1989)

$$m_{r2} = \frac{\cos \theta + 0.50572(90 - \theta + 6.07995) - 1.6364}{1}$$

(B.2)

The differences between Kasten (1966) and Kasten and Young (1989) are smaller than 0.1% for zenith angles smaller than 70°.

The relative optical air mass approximations are given at standard pressure (sea level $p_0=1013.25$ hPa). Following formula is used to find the value $m_p$ at pressure $p$:

$$m_p = m_r \frac{p}{p_0}$$

(B.3)

B.2 Relative optical ozone mass

The relative optical ozone mass $\mu$ is taken from the operational algorithm at the LKO at Arosa (Schill, 1994)

$$\mu = \frac{(R + h)}{\sqrt{((R + h)^2 - (R + r)^2 \sin^2 \theta)}}$$

(B.4)

where $R = 6.37 \times 10^6$ m is the earth radius, $r = 1800$ m is the altitude of the station Arosa, $h = 21000$ m is the altitude of the layer, and $\theta$ is the zenith angle. The ozone relative optical mass $\mu$ represents the relative geometrical thickness of a layer at 21 km above the earth's surface (ozone layer).
B.3 Integrated ozone and Rayleigh optical depths

The integrated ozone optical depth $\tau_o(\lambda)$ (sea level) is calculated as

$$\tau_o(\lambda) = \sigma_o(\lambda) \times 2.68675 \times 10^{16} X$$ (B.5)

where $\sigma_o(\lambda)$ is the ozone absorption cross section at $T=226$ K (Molina and Molina, 1986) and $X$ is the total ozone amount at Arosa [DU]. Note that 226 K corresponds to a temperature at about 21 km in the stratosphere and that 1 DU = $2.68675 \times 10^{16}$ molecules/cm$^2$.

Rayleigh optical depth $\tau_r(\lambda)$ are given at sea level e.g. by Fröhlich and Shaw (1980) and corrected by Young (1981) ($\lambda$ in [μm])

$$\tau_{r1}(\lambda) = 0.00865 \lambda^{-(3.916+0.074\lambda+0.050/\lambda)}$$ (B.6)

or by Bucholtz (1995) for a midlatitude summer model ($\lambda$ in [μm])

$$\tau_{r2}(\lambda) = 0.00651949 \lambda^{-(3.55212+1.35579\lambda+0.11563/\lambda)}$$ (B.7)
Appendix C

Mathematical and statistical methods

The statistical package "S-PLUS" (version 3.3 Release 1 for Sun SPARC) installed on Unix was used for most of the calculations presented in this report.

C.1 Multiple regression and estimation of the coefficients

All models presented in this work contain coefficients significantly different from 0 at a level of 5% (i.e. with the assumption of Gauss-Normal distributions, the error of type I is smaller than 5%\(^1\)). Variable selection was performed with a stepwise method to get subsets of significant regressors. In each estimation, we tested the error to avoid any dependence on the predictors, as well as problem of non-normality.

C.1.1 Clear-sky model

The non-linear model Eq. 5.1 is transformed into a standard multiple linear regression by taking logs.

\[
\ln \left( \frac{UV_{Bio}(X, \theta, d_c)}{d_c \cos \theta} \right) = \log(a_0) + a_1 X \mu + a_2 m \\
+ a_3 m^2 + a_4 (X \mu)^2 + a_5 X \mu m + \epsilon 
\]  

(C.1)

where \(\epsilon\) is the error and all other symbols are as described in chapter 5.

By using this form, we assume that the original model has an error that appears in the multiplicative form. The parameters \(a_\theta = \log(a_0), a_1, ..., a_5\)

\(^1\)The error of type I is the probability to reject the null hypothesis \(H_0\) if \(H_0\) is true.
and their standard deviations are estimated with the least-squares method (functions `lsfit()` or `lm()`).

### C.1.2 Combination of models

The clear-sky models are used to estimate the radiation amplificator factor and the normalisation factor to a fixed total ozone amount in chapter 5 as well as of the snow effect in chapter 7. Asymptotic confidence intervals are estimated as a function of the total ozone amount $X$ and solar zenith angle $\theta$ to determine the statistical accuracy of these values.

As an example, we present the calculation of the confidence interval for

$$RAF(X, \theta) = 100 \times (e^{a_1(0.99-1)X\mu+a_4(0.99^2-1)(X\mu)^2+a_5(0.99-1)X\mu m} - 1)$$

with the parameters $a_1$, $a_4$ and $a_5$ from Eq. C.1 and all other symbols as described in chapter 5.

With the assumption of a normally distributed error $\epsilon$ in Eq. C.1,

$$\Phi(X, \theta) = a_1(0.99 - 1)X\mu + a_4(0.99^2 - 1)(X\mu)^2 + a_5(0.99 - 1)X\mu m$$

is normally distributed and its standard deviation is given by

$$s = \sqrt{z'\Sigma z}$$

where

$$z' = (0, (0.99 - 1)X\mu, 0, 0, (0.99^2 - 1)(X\mu)^2, (0.99 - 1)X\mu m)$$

and $\Sigma$ is the variance-covariance matrix of the parameters $a_0, a_1, \ldots, a_5$ (output in `summary(lm())$corr`).

Thus the confidence interval of the estimate $R\hat{A}F(X, \theta)$ is

$$(exp(\hat{\Phi}(X, \theta) \pm 1.96 s) - 1) \times 100$$

A similar procedure is used to get the intervals for the normalisation factor to a fixed total ozone amount. The confidence intervals for the snow effect are obtained with $s$ being the sum of two terms involving the variance-covariance matrix of the parameters in the no-snow and snow models respectively.

Note that the parameterizations of the curves in chapter 7 and 8 also assumed log-normal distribution\(^2\) for estimating confidence intervals with a method similar to that described above.

\(^2\)A positive random variable $X$ is said to be log-normally distributed with two parameters $\mu$ and $\sigma^2$ if $Y = \ln X$ is normally distributed with mean $\mu$ and variance $\sigma^2$. Note that the ratio between two log-normally distributed variables is a log-normally variable.
C.2 Envelope curves

The envelopes estimated in section 6.5 are parameterized with formula Eq. C.7, which satisfies the condition that both values and first-order derivatives be identical at the beginning and the end of the year.

\[ y = k_0 + k_1 p_{s1} + k_2 p_{s2} + k_3 p_{s3} + k_4 p_{s4} + \gamma \quad (C.7) \]

with

\[
\begin{align*}
p_{s1} & = \phi - 6 \phi^5 + 5 \phi^6 \\
p_{s2} & = \phi^2 - 4 \phi^5 + 3 \phi^6 \\
p_{s3} & = \phi^3 - 3 \phi^5 + 2 \phi^6 \\
p_{s4} & = \phi^4 - 2 \phi^5 + \phi^6
\end{align*}
\]

where \( \phi = \text{julian day}/366 \) and \( \gamma \) is the error.

We propose the following automatic algorithm ("moving quantile method") to get an envelope without interactive selection of data:

1. For each day \( t \): selection of the data on days \( t - 5 \) to \( t + 5 \) with local solar true time between 11:30 and 12:30, and calculation of the 85% quantile of this data set ("moving quantile").

2. Selection of all data above the 85% moving quantile limit.

3. Estimation of the parameters \( k_0 \) to \( k_4 \) by calibrating Eq. C.7 with the data set obtained in (2). A robust regression method is applied to avoid a too large influence of extremely low values (cloudy sky).

4. The final envelope is given by the curve calculated in (3) divided by 0.925. The factor 0.925 = 0.85 + 0.15/2 is chosen such as most of the data remains under the curve.
Appendix D

Cloud variability estimator

In this appendix, we propose a measure of the cloud variability (cloud variability estimator or $cve$). This parameter aims at the following criterions:

1. $cve$ is independent on the solar zenith angle, ozone amount, aerosol and surface albedo
2. $cve$ equals zero on clear-sky days,
3. $cve$ for the global, direct and diffuse irradiance are comparable,
4. $cve$ differentiates significantly between different cloud types.

The cloud variability estimator $cve$ at time $t_0$ is defined by using irradiance values $UV$ in the time interval $[t_0 - \Delta t, t_0 + \Delta t]$ ($\Delta t$ fixed). In the first step, the series $\text{sign}(UV(t) - UV(t - 1))$ is calculated for $t \subset [t_0 - \Delta t + 1, t_0 + \Delta t]$, where $\text{sign}(x) = 1$ if $x > 0$, $\text{sign}(x) = -1$ if $x < 0$. Only relative differences larger than 1% are taken into account. In the second step, the number $S(t_0)$ of sequences of “1” and “-1” is determined. The cloud variability estimator is defined by $cve(t_0) = S(t_0)/2\Delta t$.

The $cve$ value can be calculated for the global ($cve_{\text{glo}}$), direct ($cve_{\text{dir}}$) and diffuse ($cve_{\text{dif}}$) irradiance. The $cve$ estimator is non-parametric. This procedure avoids assumptions on the clear-sky conditions (e.g. clear-sky values determined with an empirical or physical model).

As an example, the $cve$ values are given on Fig. D.1 for 4 days at Payerne (UV-Biometer measurements on Fig. 8.1 and cloud observations on Tab. 8.1). We recognize the clear-sky day for the low $cve$ values on April 7, 1997; the increasing variability of the direct irradiance ($cve_{\text{dir}}$) with the increasing cirrus coverage on November 9, 1995; the large $cve$ values on July 23, 1995 with scattered clouds and the influence of cumulus clouds on $cve_{\text{dir}}$ on August 23, 1996.

A further analysis is required to determine clusters and discrimination functions to define the relation between variability and cloud type.

\begin{table}[h]
\centering
\begin{tabular}{|c|c|c|c|c|c|c|c|}
\hline
$t_0 - 3$ & $t_0 - 2$ & $t_0 - 1$ & $t_0$ & $t_0 + 1$ & $t_0 + 2$ & $t_0 + 3$ \\
\hline
2 & 3 & 4 & 2 & 6 & 8 & 3 \\
\hline
\end{tabular}
\end{table}

The series $\text{sign}(UV(t) - UV(t - 1)) = (+1, +1, -1, +1, +1, -1)$ and $cve(t_0) = 4/6$. 

1 Example: calculation of $cve(t_0)$ with $\Delta t = 3$ and
Figure D.1: Parameter $cve_{glo}$ (left), $cve_{dir}$ (middle) and $cve_{dif}$ (right) on April 7, 1997 (clear-sky), November 9, 1995 (cirrus), July 23, 1995 (stratocumulus) and August 23, 1996 (cumulus). $\Delta t = 10$. Solar zenith angle $\leq 70^\circ$
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Leer - Vide - Empty
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• European Conference on Atmospheric UV Radiation (ECUV), 28 June - 2 July 1998, in Helsinki (Finland)
• COST meetings, 17 March 1997, in Munich (Germany), 20 - 21 October 1997, in Vienna (Austria) and, 2 - 3 July 1998, in Helsinki (Finland).