3.1.2 Climate change and radiation-induced (actinic) effects

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Climate change and actinic effects: *Anthropogenic emissions of gases and aerosols have changed the composition of the Earth's atmosphere. Resultant radiative forcings have altered atmospheric radiative transfer, which affects the biosphere, air chemistry and materials, and have caused climate changes on global and regional scales. Changes in solar spectral irradiance reaching the Earth's surface result in changes of actinic effects. One of the most noticeable results of solar actinic effects is human skin pigmentation. Reproduction of the ancestral genus Homo living in African savannas with high annual solar UV exposure was favored by a dark constitutive skin pigmentation that protected folic acid in the blood from actinic degradation. Migration waves of darkly pigmented hominins from 1.8 Ma and 0.1 Ma onwards from tropical Africa to higher latitudes considerably reduced their average annual UV exposure and subjected them large seasonal changes. Evolutionary adaptation to the changed climate by brighter constitutive pigmentation and development of facultative pigmentation (tanning according to UV exposure) avoided adverse effects on health and reproduction that result from a lack of vitamin D3 synthesis. In recent centuries, fast voluntary and forced migration to other climate zones, urbanization and increasing indoor work, changing diet and clothing, and new recreational habits have occurred at short time scales. Depending on personal skin type and outdoor behavior, inadequate adaptation of humans to solar actinic exposure can contribute to health problems. Changes in actinic effects can be expected to occur in the coming decades as a result of climate change the degree of which will also depend on reduction of CO_2 and trace gas emissions.*

Concept of Radiative Foring

Any actinic (photochemically and photobiologically induced) action of radiation is characterized by its action spectrum and the solar spectrum, which is markedly modified after solar radiation has passed the Earth's atmosphere. On global average, a radiative power of the sun of 340.2 W/m² reaches the upper limit of the Earth's atmosphere (Kopp & Lean 2011). An equivalent amount of radiative power is lost to space by backscattering of short-wave solar radiation (Earth's surface 7% + atmosphere 22% = 29%) and long-wave upward radiation (Earth's surface 12% + atmosphere 59% = 71%) (Trenberth et al. 2009, Wild et al. 2013). As a result, the Earth/Atmosphere system is in equilibrium with its environment. The Earth's surface absorbs an average solar power of 161 W/m². A small surface area of 117 × 117 km² receives an amount of solar energy that corresponds to the present global electric energy consumption of 1.93 × 10^{16} W × h per year. If the radiation balance is changed by an external effect, for example a change of solar radiation or changes due to parameters within the Earth's climate system (geosphere, oceans, cryosphere, biosphere and atmosphere), a new state of balanced equilibrium will result.

The climate relevance of individual anthropogenic changes can be described as Radiative Forcing (RF). It is the change of total net irradiance per unit area, i.e. the sum of shortwave solar irradiance and longwave atmospheric and terrestrial heat radiation; downward minus upward in W/m² at the boundary layer between troposphere (height range at midlatitudes from 0 to 12 km) and stratosphere (height range from 12 to 50 km). This definition of the RF implies that temperatures in the stratosphere are allowed to change due to a perturbation by the time of reaching radiation equilibrium after a few months time, while tropospheric and surface temperatures must remain constant. In recent years, an additional Effective Radiative Forcing (ERF) is derived that differs from the RF by describing the net radiative power not at the tropopause, but at the top of the atmosphere. In addition, the definition of ERF allows temperatures of the whole atmosphere and at the surface as well as atmospheric water content, clouds and surface albedo to respond to the Radiative Forcing (IPCC 2013). With the exception of aerosol effects by soot particles, differences between ERF and RF are small. The RF can be positive (heating) or negative (cooling). Spectral characteristics of absorption and scattering for atmospheric gases and aerosol have to be taken into account to derive RF and ERF.

Radiative Forcing leads to changes in atmospheric and oceanic circulation, affects the hydrologic cycle, changes temperature-dependent chemical reactions of atmospheric trace gases, and results in changes of the energy budget of the Earth's surface and atmosphere by affecting longwave radiative processes as well as latent and sensible heat fluxes. Internal interactions and feedbacks in the climate system can amplify or reduce the initial climate signal and actinic effects. The combination of the interacting processes and the resulting effect on climate can be simulated by coupled Atmosphere-Ocean Models and Earth-System Models.

Carbon dioxide and trace gases

Reference period for anthropogenic RF is usually the time period between 1750 (pre-industrial period) and the present time. The climate effect by individual components depends on radiative absorption characteristics of gases (spectral range and strengths of absorption bands), their concentrations and atmospheric residence times, and for atmospheric aerosol also their radiative scattering characteristics. Compared to other single anthropogenic components, carbon dioxide (CO_2) has the dominating direct effect on climate. As the final product of complete combustion of fossil fuels such as coal, crude oil and natural gas (97%) and from cement production, its global emission to the atmosphere amounts to 9.7 ± 0.5 Gt C per year (given for 2012, 1 Gt = 10^{12} kg) (Le Quéré et al. 2013). Fossil fuels still cover about 81% of the present global energy demand of about 16 TW (= 16 ×10^{12} W) that corresponds to about 0.0092% of solar radiation reaching the upper limit of the Earth's atmosphere. Somewhat more than half of the emitted CO_2 (30 to 80% with large inter-annual variation) accumulates in the atmosphere. Oceans and land surfaces (vegetation, soil) take up the remaining part of CO_2. The atmospheric CO_2 concentration increased from 275–285 ppm (1 ppm = 10^{-6} volume fraction) in 1750 to 397 ppm in 2013, which corresponds to an increase by about 42%. The present atmospheric CO_2 concentration is the result of a total emission of 240 billion tons C that accumulated in the atmosphere since 1750 by combustion of fossil fuel and anthropogenic land use (IPCC 2013). At the same time, CO_2-uptake by the oceans has resulted in acidification of ocean surface water indicated by an increase in pH by 0.1. CO_2 emitted since 1750 has an RF of 1.68 ± 0.35 W/m² and contributes approximately ¾ to the total anthropogenic ERF of 2.29 W/m² (Fig. 1).

Atmospheric concentrations of chemically reactive greenhouse gases did also increase on the long-term. As an example, concentrations of methane CH_4 (from agriculture and livestock farming, natural gas production, coal mining, garbage dumps and sewage plants) increased from 715 ppb to 1803 ppb, 1 ppb = 10^{-9} volume fraction), concentrations of fluoro-chlorocarbons (from propellants, plastics, fire-extinguishing agents, solvents and detergents) with partly long atmospheric residence times increased, and concentrations of dinitrogen mon-

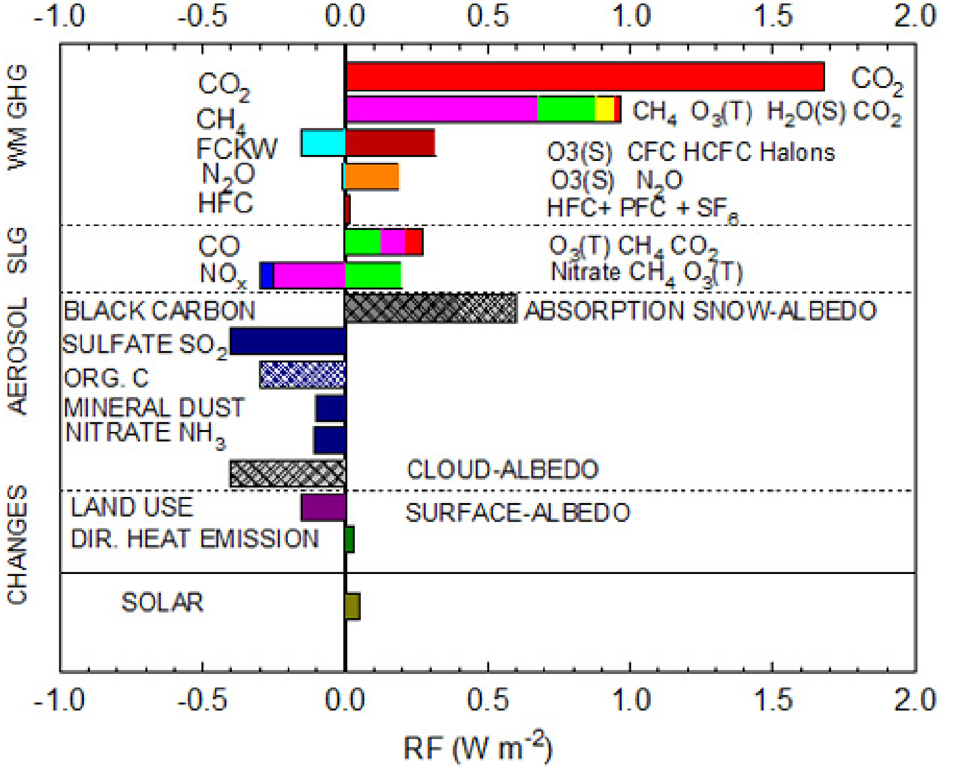

Fig. 1: Radiative Forcing in W/m² (RF or ERF) by anthropogenic greenhouse gases, aerosols and land use in 2011 referred to 1750. The direct heat emission and change of solar radiation are given for comparison (after IPCC 2013).

oxide N_2O (from livestock, legumes farming and fossil fuel combustion) increased from 270 to 319 ppb. Those trace gases do not only absorb radiation themselves, but they contribute to production and reduction of other greenhouse gases and to production of aerosols. An example is the ozone (O_3) production in the troposphere by decomposition of CH_4 and other volatile organic compounds in areas with sufficient concentrations of nitrogen oxides (NO_X). Despite its short residence time of about 22 ± 2 days (Young et al. 2013), ozone essentially contributes to tropospheric heating with an RF of (0.35 ± 0.20) W/m² and affects the air quality near the Earth's surface (more on that in chapter 3.1-3 - Mücke). Model calculations show that, on a global average, the tropospheric ozone concentration increased by 29% between 1850 and 2000 (Young et al. 2013).

Fluorochlorocarbons and dinitrogen monoxide (N_2O) are not only greenhouse gases themselves with atmospheric residence times of decades to centuries, but solar UV radiation (200–400 nm) photolyses them to radicals that effectively contribute to stratospheric ozone depletion. As a result of reduced stratospheric ozone concentration, less solar radiation in the spectral regions of UV-C (200–280 nm) and UV-B (280–315 nm) is absorbed, which leads to stratospheric cooling. Even source gases with short atmospheric residence times and weak radiation absorption such as carbon monoxide CO und nitrogen oxides NO_X indirectly contribute to the greenhouse effect by photochemically producing greenhouse gases (RF positive), but they also generate nitrate (NO_3^-) by heterogeneous reactions on the surface of aerosol particles that cause a negative RF. An early contribution to reduce the anthropogenic greenhouse effect from fluorochlorocarbons was accomplished by the »Montreal-Protocol« on Substances that Deplete the Ozone Layer' in 1987 and its later refinements and supplements (London 1990, Copenhagen 1992, Vienna 1995, Montreal 1997, Beijing 1999). Despite limiting their emission, trace gases covered by the protocol reach an RF of (0.33 ± 0.03) W/m² (Fig. 1).

Aerosol and clouds

Liquid and solid particles suspended in the air, i.e. aerosols, can be emitted directly both from natural sources (mineral dust, sea salt, pollen, fungi, spores, algae, bacteria, viruses) and anthropogenic sources (fossil fuel combustion, biomass burning), and indirectly by nucleation of gases or heterogeneous condensation of gases on existing particles. The particles with cross-sections covering a large size range from about 0.001 to 100 µm do not only absorb radiation, but especially scatter incoming radiation to different directions. Depending on their optical properties, in particular the proportion of absorption referred to extinction (= scattering + absorption) of radiation, aerosols either contribute to heating or cooling (direct effect). While due to their high absorption capacity soot particles mainly lead to direct warming, and after their deposition on the earth's surface, reduce the amount of radiation reflected by snow-covered surfaces (albedo), enhance surface warming, scattering sulfate particles (SO_4^{2-}) and nitrate particles (NO_3^-), which had been built from sulfur-containing gases (SO_2, H_2S, CS_2, COS etc.) or nitrogen compounds (NH_3, NO_X), mineral dust particles and organic carbon particles cool the lower atmosphere. Aerosol containing water soluble components such as nitrate, sulfate, carboxylic acids and sea salt particles, if they occur as tiny (ca. 0.1 µm) condensation nuclei, are very efficient in cloud formation. Condensation nuclei are capable of changing their optical properties by water uptake. Their tropospheric residence time, which depends on particle mass (particle diameter), ranges from minutes to weeks.

Indirect effects triggered by the interaction between aerosol and clouds are **i)** »the cloud-albedo effect« (first indirect effect or Twomey effect, Twomey 1977) with increasing cloud-albedo by increasing the number of small cloud condensation nuclei in water clouds or ice nuclei in mixed clouds, and ice clouds with constant liquid water content, **ii)** the »cloud-lifetime effect« (second indirect effect or Albrecht effect, Albrecht 1989) by enhancing the lifetime of clouds and lowering their precipitation efficiency and **iii)** the »half-direct effect« (Hansen et al. 1997, Johnson et al. 2004) of absorbing aerosol in clouds as a result of additional absorption of radiation followed by heating, evaporation and a tendency towards cloud dissipation. The ERF value from the interaction aerosol/clouds is between -0.1 W/m² and -1.9 W/m². Contrails that are built from aerosol and water vapor emitted by aircraft jet engines in the upper troposphere, and high-level Cirrus clouds developed from them contribute very little to heating (RF= 0.05 W/m²).

Actinic effects from aerosols and clouds mainly depend on the number of particles, their optical properties, their atmospheric residence times and, in the case of clouds, also the column amount of water. Wavelength-dependent multiple scattering of radiation by aerosol and cloud particles does not only enhance diffuse solar irradiance at the cost of direct solar irradiance, but due to the longer path of radiation through the atmosphere, tropospheric trace gases such as O_3 und sulfur dioxide (SO_2) absorb more radiation. Aerosols with large optical depths up to $\delta = 1$ in the visible spectral region – corresponding to a decrease of direct irradiance to 37% – can reduce solar global irradiance (direct + diffuse) at the Earth's surface to about 60%, while clouds piling-up vertically high to the upper troposphere and having large optical depths of $\delta >> 1$ can reduce global irradiance at the surface to about 5% of typical clear-sky values.

Present climate change

Regional increases in aerosol concentration and in cloud optical depths over the Earth's land surfaces are considered to have been the main cause of a decrease of solar irradiance by 4 to 7% in Europe, North America and Asia between 1960 und 1990. That decrease has been called »Global Dimming« (Ohmura & Lang 1989). The trend inversion starting around 1990 has been called »Global Brightening« (Wild et al. 2005). Regulatory actions for environmental protection such as the reduction in emissions of trace gases and aerosols in North America und Europe since the 1970ies and 1980ies, and also the rapid decrease of industrial production in Eastern Europe at the beginning of the 1990ies may have contributed to the brightening. Large volcanic eruptions can modify anthropogenic changes of optical depths. After the eruption of Mt. Pinatubo in the Philippines in 1991, which injected about 17 Mt SO_2 into the lower stratosphere, aerosol optical depth increased from about 0.14 to 0.3 globally corresponding to an RF of about -3 W/m², and decreased again in the subsequent years. Long-term measurements over the previous one hundred years show that in spite of the cooling effect of aerosols and clouds, temperatures at the Earth's surface increased by 0.85 ± 0.20 °C on a global average (IPCC 2013). The annual average temperature in Germany increased between 1881 and 2012 by 1.2 °C, and precipitation increased by about 10 % (DWD 2013). The temperature change does not only refer to average values, but is noticeable in a decreasing occurrence of cold nights and days, an increasing number of hot days, and a more frequent occurrence of extreme weather phenomena such as heat waves, drought periods and heavy precipitation.

In compliance with the RF and the results of climate models, temperature measurements do show long-term increases in the free troposphere and decreases in the stratosphere on a global scale. Increasing tropospheric temperatures result in a higher saturation pressure of water vapor, i.e. the capacity of air to hold water increases non-linearly with temperature (Sonntag 1994). Water vapor has a climate efficiency of 50 %, which is about double of the efficiency of clouds (25 %) and of CO_2 (20 %) (Schmidt et al. 2010). A higher water content can favor formation of clouds that enhance the dominating climate effect of water vapor by increasing it for high clouds or weaken it for low clouds. Global water vapor and total tropospheric precipitable water increased between 1970 and 2000 with the consequence that downward atmospheric longwave radiation at the Earth's surface increased as well (Wild & Ohmura 2000). Results of the NASA Water Vapor Project (NVAP) show a seasonal variation and interesting connections between total water content and atmospheric phenomena such as El Niño, but in contrast to the assumed feedback in climate simulations, no significant increase in the lower atmosphere between 1988 and 2010 (vonder Haar 2012). However, water vapor in the lower stratosphere did decrease globally by about 10% between 2000 and 2009 (Solomon et al. 2010).

In addition to its important radiative effect, water vapor contributes to chemical reactions of trace gases in the troposphere and stratosphere. By reaction with excited-state oxygen atoms $O(^1D)$ that are built from photodissociation of ozone by solar UV-B radiation, H_2O contributes to production of hydroxyl radicals (OH^-) that initiate chemical degradation of most trace gases including fluorochlorocarbons

$$O_3 + h\upsilon \rightarrow O(^1D) + O_2 \qquad \lambda < 310\text{ nm}$$

$$O(^1D) + H_2O \rightarrow 2\ OH^-$$

Actinic effects for air chemistry, biosphere and materials are most relevant in the ultraviolet (UV, 200 to 400 nm) and shortwave visible region, because the radiative energy per photon increases with decreasing wavelength of radiation.

Spectral actinic effects

Actinic effects are characterized by their spectral features and by the spectrum of solar irradiance. Action spectra have been derived for some of the numerous known actions on biosphere, air chemistry and materials (e.g. McKenzie et al. 2011). Referring to humans, actinic effects occur by radiation incident on skin and eyes. An actinic effect acting via photoreceptors in the eyes is the suppression of production of the hormone melatonin, which is built in the pineal gland from serotonin, by shortwave radiation (370–620 nm, maximum in the blue region at 450 nm) of the visible region (370–780 nm, maximum in the green at 555 nm). It promotes sleepiness and contributes to the circadiane rhythm (sleep-awake cycle) (Brainard et al. 2001). The low values of solar irradiance at the Earth's surface in the winter months at mid- and high latitudes are considered as one of the possible causes for seasonal affective disorders (winter depression). The relative contribution of spectral irradiance causing melatonin suppression to visible radiation during the diurnal cycle is up to 60 % higher at low sun angles, i.e. during the two hours closest to sunrise and sunset, than with higher sun angles. This »blue portion« is on average 20 to 40 % higher with cloudy skies than with clear sky (Feister & Franke 2011). Long-term changes of radiation cannot only enhance or reduce absolute values of melatonin suppression, but also modify the 'blue portion' of solar radiation and thus the relative contribution of melatonin suppression to visible light.

Actions caused by radiation can be perceived as unfavorable or beneficial. Examples of adverse effects of UV radiation for humans are i) acute effects such as erythema (skin reddening, sunburn with edema), pigmentation, phototoxic and photoallergic reactions, and ii) chronic, irreversible skin damage such as pore enlargement, pimples (comedones), vascular dilation (teleangiectasia) and damage of connective tissue (atrophy, solar elastosis), DNA (DesoxyriboNucleic Acid) damage, mutation and apoptosis, immunosuppression up to photocarcinogenesis (in particular non-malignant melanoma such as basal cell and squamous cell carcinoma), and eye irritation and disease (photoconjunctivitis, photokeratitis, cataracts) (Lucas et al. 2006, Brenner & Hearing 2008). UV radiation also acts on plants and animals, affects aquatic ecosystems, and impairs the properties of plastics (cf. chapter 2.5 - Häder).

Beneficial actions of UV radiation are biological mechanisms of protection and repair, and the synthesis of 7-Dehydrocholesterol to provitamin D3 in the epidermis (outer skin layer), which is important for metabolism and which is converted to vitamin D3 (cholecalciferol) by thermal isomerisation. It is transported in the blood plasma by a vitamin D binding protein from the skin cell membranes to the liver and hydroxilated there to calcidiol (25-hydroxyvitamin D = 25(OH)D) with a half–life time in the blood of 1 to 2 months. Vitamin D3 plays an important role in regulating the calcium level in the blood and for bone mineralisation (Web & Holick 1988, Holick 1992, WHO 1994, UNEP 1998). Lack of vitamin D3 can result in bone disease such as rickets with children and osteoporosis, osteomalacia, and muscle pain with adults. According to more recent findings, it is a risk factor for autoimmune diseases such as multiple sclerosis and Crohn's disease, skin diseases like psoriasis, infectious diseases like tuberculosis, and high blood pressure.

A low vitamin-D3 level of less than 20 - 40 ng/ml (50 – 100 nmol/l) 25(OH)D in the blood can be due to an insufficient UV-B-radiation exposure. During the months from November to February, solar erythemal exposure incident on the horizontal surface in low areas in Germany is between about 1 and 9 SED per day (Standard Erythemal Dose, 1 SED = $(100\ J/m^2)_{ER}$) corresponding to 0.25–2.25 MED (Minimum Erythemal Dose or minimum skin reddening with skin type 2) (Feister et al. 2011). Due to shading of the sky by buildings and trees, non-horizontal, inclined or vertical receiver surfaces, reduced time spent outdoors depending on occupational activity and recreational habit, and the small part of body skin surface not covered by clothing (hands 2 %, head 9 %), the actual daily erythemal exposure in winter may be much less than 1 SED. As the normalized action spectrum of vitamin-D3 synthesis takes slightly higher values in the UV-B region than the erythemal action spectrum, ratios between vitamin-D3 effective and erythemal radiation (VD3/ERY) depend on solar elevation angle above the horizon and on atmospheric conditions, in particular on atmospheric column ozone. Those ratios are between about 0.3 for large ozone values and low sun angles (»long path« of solar radiation) and 2 for small ozone values and high sun angles (»short radiation path«). Due to enhanced path lengths and strong ozone absorption in the lower atmosphere, optically very thick clouds such as Cumulonimbus can reduce the ratios VD3/ERY to half of the values that would have occurred without the cloud (Feister et al. 2011).

Recent research in anthropology, human genetics and paleoclimatology of the quaternary period (2.6 Ma) has essentially extended our knowledge on how humans migrated from Africa to other parts of the World, and how they adapted to new environments. There have been more and more indications that the vitamin-D3 synthesis by solar UV radiation indirectly affected skin pigmentation in the evolution of humans (Chaplin & Jablonski 2009; Jablonski 2004, 2012; Jablonski & Chaplin 2000, 2012; Juzeniene et al. 2009). The annual erythemal exposure in the tropical belt of the African continent between 30°N and 30°S amounts to about a threefold of radiation exposure reaching the Earth's surface at 45°N and S. Loss of body hair of hominins living in open grasslands and savannas more than 1 Ma ago improved their thermoregulation by sweating. Protection of the skin required against strong UV radiation was then provided by more effective skin pigmentation from Eumelanin of the skin (Brenner & Hearing 2008). Absorption of UV radiation by melanin prevents photodegradation of folic acid (Branda & Eaton 1978, Off et al. 2005, Juzeniene et al. 2013), which favors cell proliferation, and reduces the risk of neural tube defects of descendants. About 1.8 Ma ago Homo erectus and about 0.1 Ma ago *Homo sapiens* migrated northward from East Africa. At higher latitudes with lower annual UV exposure, their heavy skin pigmentation impeded a sufficient vitamin-D3 production, if it was not compensated by traditional vitamin-D rich nutrition as was common for Arctic Inuit and Ypik from raw fat of whales, walruses and seals. People with slight constitutive pigmentation and the ability to adapt to seasonal changes of solar radiation (facultative pigmentation or tanning in dependence on skin type) were able to use the lower level of vitamin-D3 effective radiation to keep a better health status that together with other influences favored their reproduction. Clothing required as cold protection at higher latitudes in winter (from about 0.083 to 0.17 Ma onward) (Toups et al. 2011) reduced the area of skin exposed to the sun. Body hair, type of activity (hunting, from about 0.01 Ma onwards agriculture) and the time spent outdoors further reduced skin exposure to solar radiation.

Forced or voluntary migration due to violence, climate change and lack of food occurred at much shorter time scales during the recent 500 years. About 11 million Africans were deported to the new World as part of the slave trade between the 16th and 19th century. Referred to the total number of 500 million people in the World in 1500, a percentage of up to 2% of the World's population moved to higher latitudes. The discovery of other continents (the Americas, Australia) by Europeans in the 15th and 16th century and their subsequent appropriation had resulted in the migration of more and more bright-skinned settlers to tropical regions. It is still being debated to what extent humans could adapt to another radiation climate within 20 to 25 generations. As a consequence of industrialization and increasing urbanization from the 19th century onwards, a growing portion of modern society lives and works inside buildings in cities where shading by structures and air pollution further reduce personal solar radiation exposure. The adverse effects of lack of radiation at high latitudes may be aggravated by inappropriate nutrition and clothing styles dictated by fashion. However, the recognition of being not adapted reveals new opportunities for each person to deliberately change attitudes.

Changes of actinic effects

Solar UV irradiance reaching the Earth' surface depends on local sun angle, i.e. season, time of the day, height above sea level, the amount and optical depths of clouds, concentration and optical properties of aerosols, surface albedo (in particular snow and ice cover), and for the shortwave UV region also the amount of ozone in the air column above the location and to a smaller extent on the concentrations of other trace gases absorbing in the UV region. As a result of fluorochlorocarbon emissions, stratospheric ozone decreased since the 1970ies. Due to the ozone decrease, UV-B radiation at the Earth's surface increased. Erythemal irradiance increased at European stations from 1980 to 2006 by 0.3 to 0.6 % per year on average (den Outer et al. 2010). About 2/3 of that increase is ascribed to decreasing cloudiness and 1/3 of it to the ozone decrease. A trend reversal from increasing annual UV-B exposure to decreasing UV-B values observed around the middle of the first decade of the present century as a consequence of increasing ozone values has occurred somewhat delayed due to decreasing aerosol concentration over that time period (Zerefos et al. 2011).

Future expected worldwide changes of erythemal radiation were calculated by a radiative transfer model using input values of ozone and cloudiness that were derived by 14 chemistry-climate models for the period from 1960 to 2100. Its results show long-term decreases of erythemal irradiation at mid- and high latitudes in both hemispheres between -3 and -16 % compared to values for the period 1975–1984 (Bais et al. 2011). Only in the tropics, where erythemal exposure is highest anyway, and that are extending to higher latitudes as a result of climate change, erythemal exposure does increase on the long-term by about 1% on average, and regionally in spring and summer up to +10% due to intensification of atmospheric circulation and decreasing cloudiness (Fig. 2). Not taken into account in this prediction are chemical processes in the lower atmosphere (troposphere), non-predictable volcanic eruptions and changes of extraterrestrial solar radiation.

Variations of radiation emitted by the sun in the shortwave UV region can be much larger than and even be reversed to changes at longer wavelengths (Haigh et al. 2010). Reconstructions of solar spectral radiation from variations of solar activity over the past 400 years showed only a small increase of total radiation over the whole spectral region by 1.25 W/m² (0.02 %), but did show higher increases in the UV by about 1% at UV wavelengths from 250 to 350 nm of up to 50 % at very short wavelengths of 121.6 nm (Krivova et al. 2010). As a result of enhanced solar activity, UV-C radiation reaching the upper and middle stratosphere increased over that period, stratospheric ozone also increased and UV-B radiation at the Earth's surface decreased globally by 9 to 13 % (Rozema et al. 2002).

Regional aspects

Further climate change will depend on future anthropogenic emissions. According to UN predictions, the World population of about 7.2 billion people in 2013 will increase to 9 billion by 2050. More than 50% of people live in cities today, of which 28 are mega cities with more than 10 million inhabitants. According to UN estimates, around 2/3 of all people will live in cities by the year 2050. The peculiar climate of cities being ‚heat islands' with reduced air exchange to their surroundings, modified fluxes of sensible and latent heat due to impervious surfaces and urban developments, enhanced concentrations of air pollutants and reduced solar irradiance, but enhanced longwave heating due to densely built-up areas, can be notably different from predicted regional climate. Actinic effects in industrialized regions and cities can thus considerably differ from regional and global effects. Direct heat emission from energy conversion and human metabolism (about 70 to 100 W per person), which is 0.027 W/m² on global average, is still small and contributes only 1% to the total RF. It does reach between 20 and 70 W/m² in urban settlements, and takes even higher values in mega cities (Shanghai 173 W/m²) (Crutzen 2004). Its contribution exceeds other anthropogenic effects in those areas by one magnitude (cf. Fig. 1). An enhanced demand in air conditioning in a hotter climate will further raise direct heat emission. Continuation and improvement of measurements as well as global and regional climate predictions of solar spectral radiation will help to make efficient use of beneficial actinic effects for maintenance and improvement of health, and effectively reduce risks of their adverse effects.

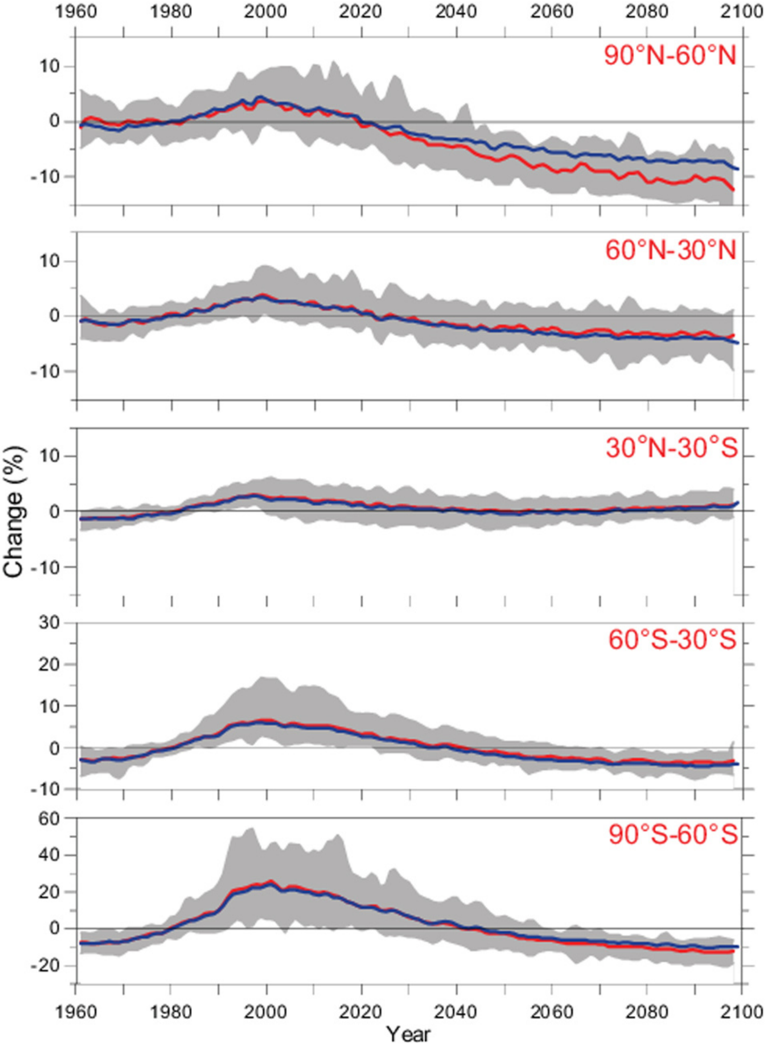

Fig. 2: *Percentage change of annual erythemal irradiation in the period 1960 to 2100 referred to the period 1975 to 1985 for geographic latitudinal belts according to radiative transfer model calculations with input data from predictions of 14 chemistry-climate models (Bais et al. 2011).*

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