Surface Radiation, Cloud Forcing and Greenhouse Effect in the Alps

A dissertation submitted to the
Swiss Federal Institut of Technology Zurich

for the degree of
DOCTOR OF NATURAL SCIENCES

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Zurich, March 2000
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Acknowledgements

I would like to thank all the people whose inspiration, encouragement and friendship allowed the realization of this PhD work. Very special thanks go to:

My advisor, Prof. Atsumu Ohmura, Institute for Climate Research ETH Zurich, who offered me the opportunity to start with the dissertation. Despite his busy schedules he always found time to answer my questions and to give me valuable advice during my trips to Zurich.

My supervisor, Dr. Rolf Philipona, PMOD/WRC Davos, who answered a thousand small and large questions. Thanks to his immense knowledge about instruments, calibration and measurements he was able to offer very valuable support and input. He was always open to new ideas and advice. He always gave me constructive feedback for my posters, abstracts and papers. His presence and readiness for discussion have greatly contributed to the progress of this work.

My co-examiner and former director of PMOD/WRC, Dr. Claus Fröhlich, who offered me the opportunity to work at the wonderful place Davos. His critical and detailed comments have been decisive influence during the research.

My co-examiner, Dr. Alain Heimo (SMI Payerne), who always had time for questions concerning data-management and data-quality of the BSRN-station Payerne.

Many people at the Swiss Institute for Snow and Avalanche Research at Davos. In particular Martin Hiller and Franz Herzog who helped with the deployment of the station “SLF Versuchsfeld” and with logger-problems.

Dr. Marcia Phillips for carefully reading and correcting my English despite the many tasks of her own research.

The Swiss Meteorological Institute (SMI), in particular the aerological station Payerne (Dr. Bruno Högger), who supported the project with instruments, dataloggers and manpower during installations and still helps with maintenance. Also to Evio Tognini from SMI Locarno-Monti, who helped deploying the ASRB-stations on the south side of the Alps and still maintains these stations. And finally, to the people at SMI Zurich for providing meteorological data and horizon values of the different stations.

All collaborators of the PMOD/WRC. The shared discussions and computation problems in all their aspects were very helpful. The shared time after work (mountain biking, jogging, cross country skiing, parties) was a good balance to the sometimes tiring work at the computer.

And last, but certainly not least, all the people who can not be listed here but helped me in many ways.

The work was financed by the Swiss National Science Foundation (Grant No. 4031-033356), the Swiss Federal Institute of Technology (ETH), Zurich (Grant No. 41-2711.5) and PMOD/WRC.
Abstract

The surface radiation budget is one of the most important components of our climate system. Every change in the radiation budget - even small in appearance - may have large consequences on the evolution of the earth-atmosphere system. These changes can either have natural causes (e.g., solar irradiance variability) or anthropogenic causes (e.g., increased greenhouse effect by emission of gases).

Surface radiation budget strongly depends on coverage of the surface (albedo) and on altitude - a fact which influences the dynamics of the atmosphere in regions with mountain chains like the Alps. The strong altitude dependence stems from the contribution of the water vapor to the overall greenhouse effect of the atmosphere. The water vapor content decreases with decreasing air temperature and thus with increasing height. The alpine region is therefore an ideal test bed for the investigation of the water vapor feedback and its impact in a changing greenhouse.

To investigate the surface radiation budget and in particular the longwave downward radiation in the Alps, the Alpine Surface Radiation Budget (ASRB) network was initiated in 1994. Meanwhile, a total of eleven stations, located between 400 and 3600 m a.s.l. in the Swiss Alps, measure short- and longwave downward radiation, air temperature and humidity with high temporal resolution and reliability. Reflected shortwave and emitted longwave radiation are measured at four expanded stations. These measurements were used to estimate the upward fluxes at the other stations, where they could not be performed for technical reasons.

A new correction method for the direct sun influence on pyrgeometers allows to measure longwave downward radiation without permanent shading. This method and a specially designed ventilation and heating system guarantee accurate measurements of short- and longwave radiation fluxes despite the harsh alpine conditions.

Four years of data of the ASRB-network allowed to quantify yearly and seasonal means of the different surface fluxes. Regional differences and altitude gradients of the surface fluxes could also be revealed with the help of the ASRB database. Yearly means of net radiation yielded positive values at all stations. The absolute values vary from about 50 W m⁻² at the lowest stations to almost zero at the highest mountain stations. This corresponds to a gradient of -1.4 W m⁻² / 100 m.

A new algorithm was developed to automatically detect clear-sky conditions with the help of longwave downward radiation, temperature and humidity. This distinction between clear-and cloudy-sky conditions allowed for the first time to investigate and determine cloud forcing for the alpine region. The data yielded a slightly negative net cloud forcing below 2000 m a.s.l., but a positive net cloud forcing above 2000 m a.s.l. This analysis of the cloud forcing allowed to quantify the negative (domination of shortwave reflection back to space) and positive (domination of longwave re-emission to the earth) impact of the cloud coverage for different altitudes under present climate conditions.

The greenhouse effect can not be directly measured from the earth surface, but longwave downward radiation is in linear relation (R² = 0.98) to the greenhouse flux. The relation between the greenhouse flux and the longwave flux has been investigated with the radiative transfer model Modtran. Accurate measurement of longwave downward radiation and cloud forcing allowed for the first time to determine yearly and seasonal means of the greenhouse flux at all ASRB-stations. Lower greenhouse fluxes were found at higher stations following an altitude gradient of -1.1 W m⁻²/100 m. A detailed analysis of clear-sky measurements and model calculations demonstrates that the decrease of longwave downward radiation and greenhouse effect with altitude is mainly driven by the decreasing water vapor content. The correlation between greenhouse flux and measured longwave downward radiation showed that a change in the greenhouse effect amplifies longwave downward radiation by a factor of 2.5.
Zusammenfassung


Der Oberflächenstrahlungshaushalt hängt stark von der Bodenbedeckung, d.h. der Albedo und der Höhe über Meer ab - eine Tatsache, welche die Dynamik der Atmosphäre in Gebirgsregionen wie die Alpen beeinflusst. Die starke Höhenabhängigkeit ist durch die Verteilung des Wasserdampfes und den damit verknüpften Treibhauseffekt bedingt. Der absolute Wasserdampfgehalt nimmt mit abnehmender Temperatur und darum mit zunehmender Höhe ab. Der Alpenraum ist darum ein ideales Testgebiet, um die Wasserdampfrückkopplung und ihren Einfluss in dem sich verändernden Treibhaussystem Erde-Atmosphäre zu untersuchen.


Vier Jahre Daten aus dem ASRB-Netzwerk ermöglichten es, jährliche und saisonale Mittelwerte der verschiedenen Strahlungsflüsse für jede Station zu quantifizieren. Die regionalen Unterschiede und die Höhengradienten der Strahlungsflüsse konnten mit Hilfe dieser Datenbasis ebenfalls zuverlässig bestimmt werden. Es zeigte sich, dass alle Stationen bis hinauf auf knapp 3600 m ü. M. positive Jahresmittelwerte der Nettostrahlung aufweisen. Die gefundenen absoluten Werte bewegen sich in einem Bereich von 50 W m⁻² an den niedrigen Stationen bis zu fast Null an den höchsten Stationen. Dies entspricht einem Gradienten von -1.4 W m⁻²/100 m.


Der Treibhauseffekt kann nicht direkt von der Erdoberfläche aus gemessen werden. Aber es besteht eine eindeutige lineare Beziehung zur langwelligen Einstrahlung mit \( R^2=0.98 \). Diese Beziehung zwischen dem Treibhauseffekt und der langwelligen Einstrahlung wurde mit dem Strahlungstransfermo-
dell Modtran untersucht. Die genaue Messung der langwelligen Einstrahlung und des Wolkeneffektes ermöglichten es damit erstmals, jährliche und saisonale Mittelwerte des Treibhauseffektes an allen ASRB-Stationen zu bestimmen. An den höheren Stationen wurden kleinere jährliche Treibhaus-Werte gefunden, was zu einem Höhengradient des Treibhauseffektes von -1.1 W m⁻²/100 m führt. Eine genaue Untersuchung der Messungen und der Modellberechnungen zeigte, dass diese Abnahme des Treibhauseffektes und der langwelligen Einstrahlung hauptsächlich durch die Abnahme des Wasserdampfgehaltes bedingt ist.
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1 Introduction

1.1 Surface Radiation Budget

The Sun is the only climate relevant energy source for our planet. Its energy is transmitted by electromagnetic radiation, which human beings, animals and plants receive in form of light and heat. Due to the earth’s position (in the ecliptic) and the movement of the earth relative to the sun (elliptic path and rotation), the distribution of solar radiation is variable with respect to latitude and season. These geographic and temporal differences are the driving motor for the general circulations within the atmosphere and the oceans. Together with the earth’s surface cover and topographic variations it is also the cause for the weather occurrence and on a longterm basis responsible for the different climates on our planet.

The solar energy and its conversion in the earth-atmosphere system determine the radiation budget at every point on Earth’s surface. On the global scale and over long periods of time the earth leaving or outgoing thermal radiation into space is equal to the absorbed solar radiation. If we look at a specific point on earth, there is an imbalance. The difference is defined as the radiation budget, which is driving latent and sensible heat fluxes between surface and atmosphere at this particular point. This budget consists of:

- the incoming solar radiation,
- the reflected radiation at the surface,
- the emitted thermal radiation from the surface,
- the emitted thermal radiation of the atmosphere.

The knowledge of the radiation budget and its components at any location is of major interest for economy (food production, solar energy use, tourism) and there is also growing importance due to a possible climate change.

1.2 Climate Change

Global climate change has attracted much scientific and public attention in recent years, as result of fears that man’s economic activities are leading to a global rise of temperature. Predictions on the basis of General Circulation Models (GCMs) presented by the Intergovernmental Panel on Climate Change (IPCC) indicate that the lower atmosphere could warm on average by about 1.5 - 4.5°C by the end of this century.

The present global mean surface temperature of 15°C is caused by the partial re-emission of the earth’s thermal radiation in the atmosphere back to the surface. This trapping of the thermal radiation is mainly due to the radiative properties of some atmospheric gases. The comfortable temperature of 15°C is approximately 30°C above that expected from a atmosphere free Earth. The process causing this warming is conventionally known as the “greenhouse effect”, and the gases responsible for it as greenhouse gases. The major greenhouse gases are water vapor (the largest contributor), carbon dioxide, ozone, methane, CFC’s, nitrogen compounds and other radiatively active gases. Real greenhouses function mainly by restricting convective heat losses, only partly by modifying the radiative regime within the enclosed space. Nevertheless, the term “greenhouse” - and its derivatives (greenhouse effect, greenhouses gases, etc.) - have entered current usage. They are hence used in this work for convenience.

This so-called “natural” greenhouse effect is therefore an essential part of the climatic system which maintains the surface temperature of our planet. In recent years it has become apparent that human
activity increases the concentration of greenhouse gases in the atmosphere to an extent that a small change in the radiation fluxes could be detected. Considerably larger changes are anticipated in the future and a significant climate warming may follow. Given the socio-economic impact that may be expected in connection with increasing greenhouse gases, there is no doubt that accurate monitoring of radiative fluxes is absolutely necessary.

The increasing atmospheric content of radiatively active gases plays a dominating role in the radiation budget of the atmosphere and the earth’s surface. Anticipated climate changes are so far predominately studied with climate models and analysed in terms of air temperature change at the earth surface, which ranges between 1.5 and 4.5°C at the time of doubled CO$_2$. However, the inter-model differences in regional temperature changes are even larger. The main cause for this uncertainty lies in our limited understanding of the important feedback mechanisms such as, the atmosphere/cloud - interaction and the ocean-atmosphere coupling.

1.3 Motivation

Radiative processes are of key importance for the understanding the evolution of the climate system. Due to satellite programs such as Earth Radiation Budget Experiment (ERBE), the net exchange of radiative energy between the global climate system and space is well established. Far larger uncertainties still exist in the knowledge of the distribution of this energy within the climate system, such as the partitioning of the energy absorption between the atmosphere and the surface, and between clouds and the cloud-free atmosphere. Large uncertainties are therefore found in the disposition of radiative energy within different GCMs. The general lack of comprehensive observed surface radiation fluxes still limit the validation of GCMs. Reliable surface radiation measurements are essential to determine the different radiation fluxes at the surface and in the atmosphere.

Even if mountain ranges account for a relatively small surface area of the globe, they are nonetheless an integral and important part of the climate system. Acting a physical barrier to atmospheric flows, mountains perturb synoptic conditions and are one of the mechanisms for cyclogenesis in middle latitudes. Because of their high altitude, mountain chains such as the Alps incorporate important elements of the climate system such as snow and glaciers. The Alps are of considerable interest, not only because a number of different climatological regimes converge on the region, but also because they are equipped with a dense observational network in comparison to other mountain chains. In the context of climate change, significant perturbations can be expected to the natural systems and these will inevitably have an influence on the sensitive ecology and economy of the Alps.

The typical grid-size of mathematical global climate models is far too large to investigate the potential impact of climate change in mountain regions. Smaller sized regional climate models are used to investigate the complex topography and its interaction with climate. For the scenario of doubled CO$_2$ concentrations, temperature is predicted to increase by 2 to 4 K in summer and by 0.5 to 1.5 K in winter. There are also indications that temperature increases more at higher elevations than at lower altitudes. These results have a high uncertainty, because it is very difficult to test such small scale models due to the small density of spatial climatological information. Detailed investigations showed a special deficiency of high quality radiation data at higher altitudes (Ohmura et al. 1996). Therefore, it is essential to measure the different radiation fluxes also at mountain stations and to analyze their altitude dependence.

1.4 The Study in Context

The importance of the radiation budget at the earth’s surface motivated the Joint Scientific Committee (JSC) to initiate the Surface Radiation Budget (SRB) program within the scope of the World Climate Research Program (WCRP) of the World Meteorological Organisation (WMO). Out of the SRB, the
Baseline Surface Radiation Network (BSRN) was born in 1992 with the aim to collect high quality radiation data and ancillary atmospheric data in different climate zones. The data of this network are used on the one hand to accurately monitor surface radiation fluxes and on the other hand to calibrate algorithms for a global climatology of the surface radiation budget from satellite measurements. Furthermore, the BSRN is also aimed at improving climate models. Switzerland has a leading position within this project, because firstly, the data center of the BSRN, the World Radiation Monitoring Center (WRMC) is located at Institute of Climate Research at the ETH in Zurich. Secondly, a number of important inputs to the BSRN recommendations concerning radiation measurements and calibration are provided by the Physikalisch-Meteorologisches Observatorium Davos and World Radiation Centre (PMOD/WRC). And finally, with the radiation station Payerne, Switzerland is maintaining one of the first established BSRN stations.

The present investigations are closely connected to the Alpine Surface Radiation Budget (ASRB) project, which was initiated in 1994 and consists of eleven radiation stations between 500 and 3500 m a.s.l. in the Swiss Alps. One of the main goals of this radiation network is to provide very accurate short- and longwave radiation fluxes at different altitudes over the Alps to achieve a better understanding of the elevation dependence of the surface radiation budget. Due to their large altitude range the ASRB stations are an ideal network for investigation of the altitude dependence of the water vapor feedback in a changing greenhouse scenario. An other important goal of the ASRB-project is the long-term monitoring of the longwave atmospheric radiation in order to directly monitor the increasing greenhouse effect.

Moreover the ASRB project allows to fill the gap of the impact of mountain ranges on the radiation budget, which is poorly simulated in GCMs and not well represented in satellite algorithms. From a technical point of view the BSRN can definitely profit from the improvements and experiences with new measurements techniques, which have been made within the ASRB network.

1.5 Objectives

This work intends to improve our knowledge about the radiation fluxes, the cloud impact and the greenhouse effect at different altitudes in the Alps. A basic condition to reach this goal is the exact knowledge of the surface radiation budget under different conditions. In particular the longwave part is only incompletely known, but very important for the understanding of the greenhouse effect. The main point of this study will hence be on the longwave radiation measurements. Major improvements were made in this topic during the last few years. Furthermore there are only very few high quality longwave radiation measurements available from higher altitudes.

This study is the first technical documentation about the ASRB-network and its data. Beside this groundwork for the ASRB-project, which is necessary for further investigations, the scientific objectives are as follows:

• to develop a method to correct the impact of the direct sun on the longwave measurement (Chap. 4),
• to develop an algorithm to distinguish between clear-sky and cloudy-sky conditions, which is necessary for further analysis (Chap. 5),
• to investigate the diurnal and annual cycle of the different radiation fluxes and to examine their impact on the change of the net fluxes with altitude (Chap. 6),
• to reveal the impact of the clouds by the determination of the short- and longwave cloud forcing for different seasons and altitudes, which will provide information about the cloud-radiation feedback processes (Chap. 7),
• to analyze the correlation between the longwave downward radiation, surface temperature, water vapor and greenhouse effect for different altitudes by comparing measurements with models (Chap. 8).
2 Measuring Surface Radiation Fluxes

2.1 Introduction

This chapter gives an introduction to the technique and the problems of surface radiation measurements and illustrates the steps taken in improving quality and accuracy of the data.

The fundamental climatic role of radiative processes has spurred the development of increasingly sophisticated models of radiative transfer in the earth-atmosphere system. Since the basic physics of radiative transfer is well known, this was thought to be an exercise in refinement. Therefore, it came as a great surprise when an international comparison between 30 radiation codes (used in climate models) yield differences of 30-70 W m\(^{-2}\) among longwave atmospheric fluxes (Luther et al. 1988) and of 60-100 W m\(^{-2}\) among shortwave solar fluxes (Fouquart et al. 1991). In order to resolve the discrepancies, a field program was recommended. According to three newer studies (Cess et al. 1995, Ramanathan et al. 1995, Pilewski and Valero 1995), comparing model results and observational data, clouds appear to be absorbing more incoming solar radiation than they are expected from theoretical evaluations. A recent experiment (Halthore et al. 1998) demonstrated that the models overestimate the diffuse surface irradiance even under clear-sky conditions.

All these studies show that the current understanding of atmospheric physics and radiative transfer is not complete. The cause of the disagreements and uncertainties can only be found with the help of extensive measurements of the irradiances and atmospheric components from the surface and from space.

In the late-1980s the WCRP insisted on radiative flux accuracies of at least 10 W m\(^{-2}\) for the surface and the top of the atmosphere fluxes. With an accuracy of about 1-2% this goal could almost be reached in the field of the shortwave radiation measurement, where fluxes from 0 to 1300 W m\(^{-2}\) are observed. The real problem were longwave radiation measurements. They had to rely on pyrradiometers or pyrgeometers, which had a relative accuracy of about ±20 W m\(^{-2}\) (Dutton 1993, WCRP/WMO 1991). The following sentence can be found in Ellingson and Wiscombe (1996): "Pyrgeometers, when operated routinely (rather than painstakingly nursed), have accuracy no better than 5% or 10-20 W m\(^{-2}\) and are notoriously difficult to calibrate and maintain." To fulfil the above mentioned goal of measuring the increase of longwave downward radiation due to a possible climate change, the accuracy had to be improved by at least a factor of five.

2.2 Definitions and Measurements Technique

2.2.1 Concept of Radiative Fluxes

The solar radiation spectrum is contained to more than 99.9% in wavelengths between 0.2 and 100 µm. The radiation energy of the sun is reflected and absorbed in the atmosphere and at the surface. The absorbed solar radiation is emitted as so-called terrestrial radiation. Hence there is a conational distinction between solar and terrestrial radiation relating to their sources. The first is the radiation originating from the sun, which reaches the surface as direct or diffuse (indirect) solar radiation. The second is the radiation originating (emitting) from the surface or the atmosphere and is called terrestrial or thermal radiation. The maximum energy of the solar radiation is at approximately 0.55 µm, that of the terrestrial radiation at approximately 11.0 µm. Both spectra cross each other at approximately 4.0 µm at relatively low power (Figure 2.1).
This crossing point is often used to distinguish between shortwave and longwave radiation. The definition of shortwave radiation is hence all radiation energy below 4.0 μm and longwave radiation all radiation energy above 4.0 μm. This definition is almost identical to the solar, resp. terrestrial radiation, but not exactly, because there is still a not negligible part of energy of the solar and terrestrial radiation beyond the 4.0 μm limit. Thus the terms shortwave (SW) and longwave radiation (LW) should be used synonymously to solar and terrestrial radiation, i.e. according to their source and not limited to wavelength ranges.

### 2.2.2 Principle of Radiation Measurements

Radiation can either be measured spectrally or broadband. Both techniques have their advantages and disadvantages. Due to the simplicity and high accuracy, broadband instruments are mainly used for investigating radiation budget fluxes. Generally, these instruments are used with the radiation receiver placed horizontally so that the incident radiation over the full elevation range and over an azimuth of 360 degrees is measured. Because these hemispherical-type of instruments are normally used for routine measurements (i.e. left permanently outside), they must be able to withstand the effects of all types of meteorological conditions. The detector is protected by a glass, quartz or silicon dome, which also defines the spectral range of the radiation being measured. The incoming radiation is absorbed by a black receiver (heat transducer, e.g. thermopile) causing a temperature difference, which can be measured as voltage signal proportional to the incoming radiation.

For the measurement of the radiation budget some manufactures sell “two in one”-instruments called net-radiometers, which measure the net fluxes directly. For some applications this may be practical. Many questions in climatology need the single components of the radiation budget. It is therefore
better just to use the same type of instruments (turned upside down) for the upward fluxes and calculate the radiation budget.

2.3 Shortwave Radiation

The shortwave radiation can be divided into a direct and a diffuse component. The direct radiation has to be measured with a narrow view angle instrument called a pyrheliometer. The diffuse component is defined as the indirect (scattered) solar radiation. The global radiation is the sum of the direct and diffuse component and is the most widely measured radiation flux. Both diffuse and global radiation fluxes are normally measured with an instrument called a pyranometer (PYR). To measure the diffuse radiation the pyranometer has to be shaded. The most widely used types are the Eppley Precision Spectral Pyranometer (PSP) from Eppley Laboratories and Kipp&Zonen CM21 pyranometer. The CM21 is a more recent instrument. It has a smaller non-linearity and cosine error than the PSP. The detector of both instruments is protected by a double glass dome to minimize the heat transfer between the inner dome and the thermopile. The glass dome is transparent between about 0.3 and 2.8 μm, which corresponds to about 96% of the extraterrestrial solar spectrum. The remaining energy is taken into account by calibration. The shortwave flux (SW) can easily be calculated by dividing the pyranometer signal \( U_{PYR} \) by the calibration coefficient \( C_{PYR} \).

\[
SW = \frac{U_{PYR}}{C_{PYR}},
\]

whereas \( U_{PYR} \) is normally measured in [mV], \( C_{PYR} \) in [μV / W m\(^{-2}\)] and the shortwave flux SW is expressed in [W m\(^{-2}\)].

2.3.1 Absolute Uncertainty

The shortwave reference is determined by pyrheliometers, which measure the direct solar radiation. Seven pyrheliometers constitute the so-called World Standard Group (WSG), of which the mean value realizes the World Radiometric Reference (WRR) at the World Radiation Center (WRC) in Davos. The WRR is re-determined every 5 years during an International Pyrheliometer Comparison (IPC). The calibration of pyranometers are performed with the direct WRR measurement and the diffuse measurement of a shaded reference pyranometer. The absolute accuracy of a calibrated and well maintained CM21 pyranometer is about 1%.

2.3.2 Problems

Zero Offset

Negative pyranometer signals during the night are called “zero offset” or “nighttime offset”. The cause for the negative signals is the gradient between the dome and the instrument body. The gradient is especially large during clear nights, because the dome temperature is decreased by the cool night sky whereas the instrument body stays at ambient temperature. This causes a radiative cooling of the thermopile, which produces a negative signal during the night instead of zero. The zero offset can be reduced by the use of an appropriate ventilation system, which blows ambient or slightly heated air uniformly around the dome.

A bad ventilation system, in contrast, can increase the offset by unintentionally heating the body of the instrument. The air is unintentionally warmed either by the ventilation motor itself or through a tube exposed to the sun. This causes large temperature gradients within the instrument and can produce zero offsets up to -20 W m\(^{-2}\). It is best to minimize the gradient by an appropriate ventilation system, i.e. without tubes. With a good ventilation the offset is even in an extreme climate (dry and cold atmo-
Measuring Surface Radiation Fluxes

sphere) very seldom larger than -5 W m\(^{-2}\). It is believed that at least part of the offset is calibrated in, because some temperature gradient also occurred during the calibration procedure. The more different the actual conditions are from those during calibration, the larger the error is. However, there is no general agreement on this problem.

**Calibration**

An unsolved issue is that all pyranometers are calibrated unventilated for global radiation measurement, although a number of them are later used ventilated and/or for diffuse measurements. A general problem within radiation measurements is the variable absorption of the atmosphere due to variable water vapor content. This has an impact on the part of energy outside the radiometric filters and hence not measured during operation. The glass dome of a pyranometer is transparent between about 0.3 and 2.8 \(\mu\)m, which corresponds only to a part of the solar spectrum. The remaining energy is taken into account by calibration as explained at the beginning of this section. This statement is only valid for the state of the atmosphere during calibration. In the tropics, for example, the transparency of the dome corresponds to about 99% of the local solar spectrum. This implicates that pyranometers which are calibrated in relatively dry location like Davos measure too much if they are used under more humid conditions and too little in dryer environments (Table 2.1). The uncertainty of a pyranometer calibrated under different atmospheres than it is used (this is normally the case), is therefore rather 2% than 1%.

<table>
<thead>
<tr>
<th>Direct horizontal SW flux</th>
<th>Tropical Atmosphere 0m, day 213, z 45°</th>
<th>Mid. Lat. Sum. Atm. 1500m, day 213, z 45°</th>
<th>Sub. Arct. Win. Atm. 3500m, day 60, z 45°</th>
</tr>
</thead>
<tbody>
<tr>
<td>&gt; 2.8 (\mu)m</td>
<td>649 W m(^{-2})</td>
<td>776 W m(^{-2})</td>
<td>98 W m(^{-2})</td>
</tr>
<tr>
<td>6.9 W m(^{-2})</td>
<td>11.4 W m(^{-2})</td>
<td>19.3 W m(^{-2})</td>
<td></td>
</tr>
</tbody>
</table>

**2.4 Longwave Radiation**

The longwave radiation fluxes can be measured with a pyrradiometer or a pyrgeometer. A pyrradiometer is sensitive to the radiation of wavelengths form 0.3 - 60 \(\mu\)m, whereas a pyrgeometer filters the solar radiation and is only sensitive between 3.5 - 60 \(\mu\)m. The problems with the measurement of longwave radiation are somewhat more complicated than those encountered with shortwave radiation, because the filter has to be transparent - and the sensor has to be flat - over a very wide wavelength range. To minimize direct and indirect solar effect both instruments should be shaded.

Several studies (Marty 1993, WMO 1983) have shown that pyrradiometers have some specific disadvantages:

- The sensitivity is not limited to the infrared, but includes the whole shortwave region, which has to be subtracted.
- The necessary pyranometer signal to compute the LW is an additional source of errors.
- The air-inflated polyethylene dome is weak and can not be used in unattended mode.
- It is difficult to find a “correct” calibration constant due to the wide wavelength range

As a consequence of these problems more and more pyrgeometers are used. The BSRN recommended pyrgeometers as the most accurate instrument for measuring longwave irradiance (WCRP/WMO, 1998). The most widely used pyrgeometer is the Precision Infrared Radiometer (PIR) from Eppley Laboratory. The PIR is a development from their Precision Spectral Pyranometer (PSP), but with a silicon rather than a glass dome. The following paragraphs are focused on the Eppley PIR, although most of the points are also valid for other pyrgeometers.
2.4.1 The Thermal Behavior of the Pyrgeometer

A pyrgeometer consists of a black-painted thermopile with the lower junction set in contact with a temperature reservoir in the form of a heavy metal base, and the upper junction set radiatively exposed to the atmosphere through a filter dome that rejects radiation below about 3.5 μm and insulates the thermopile from direct heat transfer by the air. The thermopile sensor surface absorbs and emits as a black-body at the temperature $T_r$. The output of the thermopile can be expressed as a voltage signal $U_{PIR}$ [mV] proportional to the temperature difference $\Delta T$ between the upper and lower junction of the thermopile. This temperature difference is maintained by thermal conduction, i.e., heat transfer from the instrument body to balance the net radiation loss $F_{net}$ from the top of the thermopile:

$$F_{net} = F_{in} - F_{out} = U_{PIR}/C_{PIR},$$

where $C_{PIR}$ [$\mu$V/W m$^{-2}$] is the sensitivity of the thermopile, $F_{in}$ the measured incoming flux, $F_{out}$ the outgoing flux just above the thermopile. The temperature $T_s$ of the upper thermopile junction can not be measured directly, but it can be derived from the body temperature $T_b$ [K], measured at the cold (lower) junction of the thermopile (Figure 2.2), as shown by Albrecht et al. (1974) and Philipona et al. (1995). $F_{out}$ can therefore be written with the help of the Stefan-Boltzmann law as

$$F_{out} = \varepsilon_s \cdot \sigma \cdot T_b^4.$$  \hspace{1cm} (2.3)

$\sigma$ is the Stefan-Boltzmann constant and $\varepsilon_s$ the emittance of the thermopile, which is normally set to unity and implicitly included in the calibration. After rearranging Equation 2.2 the longwave irradiance $LW$ [W m$^{-2}$] measured with a pyrgeometer can be expressed as

$$LW = F_{in} = U_{PIR}/C_{PIR} + \sigma T_b^4.$$  \hspace{1cm} (2.4)

The transmittance of the silicon dome with its vacuum-deposited low-pass interference filter has a cut-on at approximately 3.5 μm and the transmission varies between 0.2 and 0.4 in the region of 3.5-50 μm (Figure 2.3). The large dome absorption causes temperature differences between the dome and the body of the pyrgeometer, and hence, an additional thermal irradiance on the sensor surface, which is proportional to the difference between the dome temperature $T_d$ [K] and the body temperature $T_b$ [K]. Pyrgeometers have therefore been equipped with a thermistor mounted at the lower rim of the dome to measure the temperature of the silicon dome (Figure 2.2). Albrecht and Cox (1977) introduced an additional term in the pyrgeometer equation to correct for the exchange of radiant energy between the dome and the body in the form of:

$$LW = U_{PIR}/C_{PIR} + \sigma T_b^4 - k\sigma(T_d^4 - T_b^4),$$

where $k$ is a “correction” factor proportional to the emittance and transmittance of the dome. This is the normally used equation for determining the longwave radiation from a pyrgeometer.

2.4.2 Improvements and Modifications

Dome Temperature Measurement

On the way of improving the longwave measurements some authors (Foot 1986; Philipona et al. 1995) reported large dome temperature gradients and concluded that the dome temperature can not be measured satisfactorily at a single point. Philipona et al. (1995) therefore introduced a new dome temperature measurement to provide an improved estimate of the average dome temperature using three thermistors separated by 120° (in direction N, SE, SW) and glued at 45° elevation instead of the one thermistor at the rim (Figure 2.2).
Re-Evaluation of the Thermal Flux

Philipona et al. (1995) performed a re-evaluation of the thermal flux balance of the pyrgeometer and introduced a new equation with three correction factors $k_1$, $k_2$, $k_3$:

$$LW^\downarrow = \frac{U_{\text{PIR}}}{C_{\text{PIR}}} \left( 1 + k_1 \sigma T_B^3 \right) + k_2 \sigma T_B^4 - k_3 \sigma (T_D^4 - T_B^4)$$

(2.6)

The $k_{1,2,3}$ values comprise dome characteristics such as absorptance, transmittance and reflectance, as well as the emittance of the receiver surface of the thermopile. In comparison to the old Albrecht and Cox equation (Equation 2.5) this new approach improves the accuracy of the calculated flux by about 1-1.5 W m$^{-2}$.

![Figure 2.2: The most important parts of an Eppley PIR. On modified pyrgeometers the old dome temperature thermistor at the rim is replaced with 3 thermistors separated by 120° direction and glued at 45° elevation.](image)

A recent international Round Robin pyrgeometer calibration experiment carried out by eleven laboratories showed on the one hand notable differences of up to 20% from the median of the responsivities of five participating PIR pyrgeometers (Philipona et al. 1998). On the other hand, in six laboratories the absolute deviation around the median of the deviations of the five instruments was less than 1%. This small scatter suggests that the PIR pyrgeometers were stable at least during the two years of Round Robin tests and that it is possible to reproduce the responsivity $C$ of PIR pyrgeometer consistently and within the precision required for climate applications. The BSRN community as the initiator of the experiment recommended as a conclusion, that it is important:

- To have more than one calibration point.
- That the body temperature of the PIR during calibration should be close to the annual mean of the ambient temperature of the site at which the PIR will be deployed.
- That the blackbody temperature is ca.10-25°C below the pyrgeometer body temperature.
- That during the calibration the dome temperature must be varied individually. Otherwise the dome correction factor $k$, which has to be precisely known for field measurements, can not be determined.

Philipona et al. (1998) also reported that dome correction factors were more difficult to determine. They assumed that the main reason for the larger scatter was due to the dome temperature measure-
ments, which seems to be influenced differently by the various calibration devices and, hence, that the single dome thermistor at the rim does not indicate a representative dome temperature.

**Absolute Uncertainty**

In contrast to the shortwave measurements there is no absolute reference available for longwave measurements. A first absolute instrument was developed and built at PMOD/WRC in 1998. The approach is based on a reference black body radiation source and is called a Planck Calibrated Sky Radiometer (PCSR). It consists of a windowless pyroelectric detector with a narrow field of view. The measurements are performed by scanning the sky during a perfectly clear night. A Gaussian integration method allows to calculate the total downward radiation, which can be compared with the pyrgeometer data. First comparisons showed encouraging results. The deviations of the pyrgeometer measurements to measurements of the absolute PCSR instrument have always been within ±2 W m⁻².

**2.4.3 Problems**

One reason for this deviation to the current (not yet accepted) absolute reference may be the spectrally variable transmittance of the pyrgeometer dome between 0.15 and 0.40 (Figure 2.3), which unfortunately has a small transmittance drop in the middle (9 µm) of the atmospheric window region (7-14 µm), where about 93% of the pyrgeometer signal \( U_{\text{PIR}} \) originates from. The reason is the fact that a pyrgeometer effectively measures the difference between the black body spectrum of its temperature and the atmospheric spectrum. Thus the driving radiation for the pyrgeometer comes from the two atmospheric windows which extend over the wavelength range of 7-14 µm.

![Figure 2.3: Comparison of a mid latitude summer spectrum for Davos (1600 m a.s.l.) calculated with MODTRAN and compared to three Planck curves. The solid brown line shows a typical transmittance curve of a PIR dome. The upper percentage numbers give the origin of the measured pyrgeometer signal.](image-url)
Another problem is the larger transmittance drop at about 15 μm, where the maximum of the CO₂ emission bands is located. The sensitivity coefficient \( C_{\text{PIR}} \) which is determined during the calibration, is based on an average transmission of the calibrated PIR. The small transmittance in the middle of the atmospheric window and in the middle of the CO₂ emission bands (Figure 2.3 and Figure 2.4) may therefore be responsible for an underestimation of the irradiance at this wavelength.

An other reason is that the black body radiation of the calibration apparatus is a Planck curve, whereas the spectrum of the longwave atmospheric radiation with its atmospheric window is non-Planck. This is important for the part of the spectrum outside the transmittance of the PIR dome. The remaining energy outside the filter limits considered through the calibration is based on a Planck curve. The thus produced error depends on the state of the atmosphere and the calibration temperatures.

The problems with the variable cut-on of the interference filter of the PIR dome and its error due to direct solar radiation are discussed in Section 2.4.4 and can be ignored if the instrument is shaded.

There are also some other problematic points associated with the use of a pyrgeometer. For many projects and field campaigns it is difficult or even impossible (Chapt. 3) to shade the pyrgeometer. If the PIR is equipped with the new dome thermistors, well ventilated and deployed at sea level (i.e. small atmospheric window) the error may be tolerable. The errors are biggest if the PIR is located at a high mountain (i.e. large atmospheric window) station in summer (i.e. large solar power) due to the not negligible amount of direct solar radiation above the low cut-on of the interference filter.

2.4.4 Requirements for Accurate Pyrgeometer Measurements

Removal of Internal Battery

An Eppley PIR directly shipped from the factory has a thermistor-battery-resistance circuit incorporated to compensate for the emitted radiation. Albrecht and Cox (1977) identified systematic errors due to this electronic compensation of the thermal emission of the thermopile. The voltage uncertainty and the nonlinearities of the circuit at extreme temperatures induce this errors. This battery problem can easily be overcome by removing the battery and instead measuring the body temperature \( T_B \) at the cold junction of the thermopile and computing the outgoing radiation.

Shading

Pyrgeometers deployed to measure downward atmospheric radiation have to be operated with a shading disc, although it is not explicitly mentioned in the Eppley instruction sheet for PIRs. The shading prevents excessive heating of the dome by the sun, which is first of all important for the majority of PIR using just the one rim thermistor to measure the dome temperature.

The transmittance of the silicone dome with its vacuum-deposited low pass interference filter has a cut-on at approximately 3-4 μm and the transmittance varies between 0.2 and 0.4 in the region of 3.5-50 μm (Miskolczi and Guzzi 1993). The small missing part above 50 μm is included during calibration. Due to the extend of the solar spectrum above 3.5 μm, part of it is unfortunately also measured by the PIR, which, by definition, is part of the shortwave radiation. Unfortunately the blocking edge of the interference filter is very variable between different domes (Figure 2.4). These differences produce significant errors because the direct solar irradiance between 3-4 μm can not be ignored. The amount of shortwave radiation above the cut-on of the PIR dome depends not only on the dome but also on the atmospheric composition (i.e. mainly the concentration of water vapor) and path length (air mass). A adequate shading disc solves this problem by shielding the sensor from receiving this infrared fraction of the solar beam, which can reach up to 15 W m⁻² (Chapt. 4).
FIGURE 2.4: Transmittance spectra of three different PIR domes (data from Miskolczi and Gazzi, 1993). Unfortunately the spectrum differences between 3 and 4 μm, where the solar radiation should be blocked, can be very significant. Another problem is the transmittance drop in the middle of the atmospheric window, which is discussed in SECTION 2.4.3.

Ventilation
A radiometer dome has to be ventilated to avoid water deposits, like dew or frost, on their apices when the air temperatures decreases. For the pyrgeometer there is an additional reason to ventilate the dome exposed to the diffuse radiation; the ventilation permits also allows to reduce the solar heating of the dome, to minimize the effect of the wind on the dome and to decrease any temperature differences within a pyrgeometer. To attain all the mentioned goals it is important that the ventilation system blows a uniform air flow from below around the PIR.

2.5 Summary
The limitations of short- and longwave radiation measurements have been displayed. Significant improvements have been made in the longwave part, where the relative measurement uncertainty could be decreased from formerly ±15 W m⁻² to ±2 W m⁻² (TABLE 2.2). This was achieved by the improvement of the following three points:

- The calibration technique, which allows to change the case-, dome-, and blackbody-temperature separately.
- Three dome temperature at 45° elevation instead of only one thermistor at rim.
- The pyrgeometer formula by re-evaluation of the PIR energy balance.
First comparisons with a newly developed absolute longwave standard demonstrated that the absolute accuracy of well calibrated PIRs is within ±4 W m\(^{-2}\). The accuracy can only be improved by comparing the PIR with the absolute standard under a clear night sky. Good consistence can be achieved between the (not yet accepted) absolute reference and the PIR by carefully adjusting the sensitivity factor \(C_{\text{PIR}}\). After such a comparison the absolute uncertainty is decreased to ±2 W m\(^{-2}\). This accuracy allows to use PIRs for monitoring changes in longwave radiation due to an increasing greenhouse effect.

**TABLE 2.2: Accuracy improvements of longwave radiation measurement due to the new dome temperature measurement, the better calibration method and the new pyrgeometer formula.**

<table>
<thead>
<tr>
<th>Accuracy with... (according SECTION 2.4.2)</th>
<th>Rel. Measurement Accuracy [W m(^{-2})]</th>
</tr>
</thead>
<tbody>
<tr>
<td>Factory calibration</td>
<td>±10</td>
</tr>
<tr>
<td>Three dome thermistors and new calibration</td>
<td>±3</td>
</tr>
<tr>
<td>New formula</td>
<td>±2</td>
</tr>
</tbody>
</table>

**FIGURE 2.5: Hoar frost at ASRB-station Weissfluhjoch after a stormy night. The domes of the instruments are still free, i.e. the corresponding data need not be flagged.**
3 The ASRB-Network

"The enlargement of the longwave flux network is desirable because of the presently inadequate capability of deriving surface fluxes from satellite data even for cloudless areas." (NASA Reference Publication No. 1169, 1986).

3.1 Location and Measured Fluxes of the Stations
The main objective of the Alpine Surface Radiation Budget (ASRB) project is the analysis of short- and longwave radiation data from different altitudes with the best possible accuracy. One condition for this goal was the establishment of the ASRB-network in the Swiss Alps.

FIGURE 3.1: Geographical distribution of the eleven ASRB-stations in Switzerland. The ASRB-network is part of the Swiss Atmospheric Monitoring Network CHARM (Heimo et al. 1998), which consists of various sites measuring solar, IR, UV and spectral radiation.

Due to logistic and financial reasons the ASRB stations had to be deployed at locations where an already existing infrastructure was available. The lower stations were therefore deployed at locations where the the Swiss Meteorological Institute (SMI) maintains automatic weather stations (ANETZ). A few years ago SMI and the Swiss Institute for Snow and Avalanche Research (SLF) established some wind stations in the framework of an additional meteorological network (ENET). The location of these stations offered a ideal platform for the ASRB mountain stations, because their positions mostly had a very good horizon. The installation of the ASRB-stations started in 1994 and today the network consists of eleven remote automatic radiation stations between 370 and 3580 m a.s.l. (Figure 3.1 and Table 3.1), which all measure short- and longwave downward radiation.
The ASRB-Network

**TABLE 3.1: Type, altitude and coordinates of the eleven ASRB-stations.**

<table>
<thead>
<tr>
<th>Station</th>
<th>Type</th>
<th>Altitude</th>
<th>Geogr. Coordinates</th>
<th>Swiss Coordinates</th>
</tr>
</thead>
<tbody>
<tr>
<td>Locarno-Monti</td>
<td>PMOD/ANETZ</td>
<td>370 m</td>
<td>08 47 / 46 10</td>
<td>704 150 / 116 325</td>
</tr>
<tr>
<td>Payerne</td>
<td>PMOD/ANETZ</td>
<td>490 m</td>
<td>06 57 / 46 49</td>
<td>562 100 / 184 925</td>
</tr>
<tr>
<td>Davos See</td>
<td>PMOD</td>
<td>1550 m</td>
<td>09 51 / 46 49</td>
<td>783 788 / 187 550</td>
</tr>
<tr>
<td>Davos</td>
<td>PMOD/ANETZ</td>
<td>1610 m</td>
<td>09 51 / 46 49</td>
<td>783 588 / 187 488</td>
</tr>
<tr>
<td>Cimetta</td>
<td>PMOD/ANETZ</td>
<td>1670 m</td>
<td>08 48 / 46 12</td>
<td>704 300 / 117 525</td>
</tr>
<tr>
<td>Männlichen</td>
<td>ENET</td>
<td>2230 m</td>
<td>07 57 / 46 37</td>
<td>638 490 / 162 550</td>
</tr>
<tr>
<td>SLF-Versuchsfeld</td>
<td>PMOD/SLF</td>
<td>2540 m</td>
<td>09 49 / 46 50</td>
<td>780 865 / 189 040</td>
</tr>
<tr>
<td>Weissfluhjoch</td>
<td>ENET</td>
<td>2690 m</td>
<td>09 49 / 46 50</td>
<td>780 613 / 189 625</td>
</tr>
<tr>
<td>Weissfluhjoch_out</td>
<td>PMOD/ANETZ</td>
<td>2690 m</td>
<td>09 49 / 46 50</td>
<td>780 613 / 189 625</td>
</tr>
<tr>
<td>Eggishorn</td>
<td>ENET</td>
<td>2895 m</td>
<td>08 06 / 46 26</td>
<td>650 290 / 141 900</td>
</tr>
<tr>
<td>Les Diablerets</td>
<td>ENET</td>
<td>2965 m</td>
<td>07 12 / 46 20</td>
<td>581 913 / 130 613</td>
</tr>
<tr>
<td>Gornergrat</td>
<td>ENET</td>
<td>3110 m</td>
<td>07 47 / 45 59</td>
<td>626 825 / 092 465</td>
</tr>
<tr>
<td>Jungfraujoch</td>
<td>PMOD/ANETZ</td>
<td>3580 m</td>
<td>07 59 / 46 33</td>
<td>641 775 / 155 275</td>
</tr>
</tbody>
</table>

**TABLE 3.2: Measured elements and station identification number (ID).**

<table>
<thead>
<tr>
<th>Station</th>
<th>Abbreviation</th>
<th>Radiation Fluxes</th>
<th>Meteo Elements</th>
<th>ID</th>
</tr>
</thead>
<tbody>
<tr>
<td>Locarno-Monti</td>
<td>LOM</td>
<td>SW↓, LW↓</td>
<td>P, T, U</td>
<td>409</td>
</tr>
<tr>
<td>Payerne</td>
<td>PAY</td>
<td>SW↓, LW↓</td>
<td>P, T, U</td>
<td>410</td>
</tr>
<tr>
<td>Davos See</td>
<td>SEE</td>
<td>SW↑, LW↑</td>
<td>T, U</td>
<td>412</td>
</tr>
<tr>
<td>Davos</td>
<td>DAV</td>
<td>SW↓, LW↓, shaded LW↓</td>
<td>P, T, U</td>
<td>401</td>
</tr>
<tr>
<td>Cimetta</td>
<td>CIM</td>
<td>SW↓, LW↓</td>
<td>P, T, U</td>
<td>403</td>
</tr>
<tr>
<td>Männlichen</td>
<td>MAE</td>
<td>SW↓, LW↓</td>
<td>T, U</td>
<td>406</td>
</tr>
<tr>
<td>SLF Versuchsfeld</td>
<td>VSF</td>
<td>SW↓, LW↓, SW↑, LW↑</td>
<td>T, U</td>
<td>411</td>
</tr>
<tr>
<td>Weissfluhjoch</td>
<td>WFJ</td>
<td>SW↓, LW↓</td>
<td>P, T, U</td>
<td>402</td>
</tr>
<tr>
<td>Weissfluhjoch_out</td>
<td>WFO</td>
<td>SW↑, LW↑ (temporarily)</td>
<td>P, T, U</td>
<td>422</td>
</tr>
<tr>
<td>Eggishorn</td>
<td>EGH</td>
<td>SW↓, LW↓</td>
<td>T, U</td>
<td>404</td>
</tr>
<tr>
<td>Les Diablerets</td>
<td>DIA</td>
<td>SW↓, LW↓</td>
<td>T, U</td>
<td>408</td>
</tr>
<tr>
<td>Gornergrat</td>
<td>GOR</td>
<td>SW↓, LW↓</td>
<td>T, U</td>
<td>407</td>
</tr>
<tr>
<td>Jungfraujoch</td>
<td>JFJ</td>
<td>SW↓, LW↓</td>
<td>P, T, U</td>
<td>405</td>
</tr>
</tbody>
</table>

The station Payerne was installed to connect the alpine measurements to the Swiss "Mittelland" and due to the extended measurements of the BSRN-station. The station Locarno-Monti represents together with Cimetta and Gornergrat the southern side of the Alps. The stations Davos, SLF Versuchsfeld and (only temporary) Weissfluhjoch (2690 m) additionally measure the short- and longwave upward fluxes for investigations of the surface radiation budget at different altitudes. For auxiliary information the standard meteorological elements temperature (T), humidity (U) are also measured at the eleven stations; pressure (P) is measured at the seven ANETZ-stations (Table 3.2).
### Table 3.3: Location, surroundings and ground type of the ASRB-stations.

<table>
<thead>
<tr>
<th>Station</th>
<th>Location and environment</th>
</tr>
</thead>
<tbody>
<tr>
<td>Locarno-Monti</td>
<td>Roof of SMI building on valley slope, surrounded by houses and trees</td>
</tr>
<tr>
<td>Payerne</td>
<td>Research field SMI, grassland with crops in the vicinity</td>
</tr>
<tr>
<td>Davos See</td>
<td>Wind mast of SMI on valley floor with grassland underneath</td>
</tr>
<tr>
<td>Davos</td>
<td>Roof of PMOD building on valley slope, surrounded by grassland</td>
</tr>
<tr>
<td>Cimetta</td>
<td>Final mast of ski lift, local peak surrounded by grassland</td>
</tr>
<tr>
<td>Männlichen</td>
<td>Roof of cable-car mountain station, located on a grass ridge</td>
</tr>
<tr>
<td>SLF Versuchsfeld</td>
<td>Research field of SLF, located in small valley with rocky ground</td>
</tr>
<tr>
<td>Weissfluhjoch</td>
<td>Wind mast of SLF on rocky mountain peak</td>
</tr>
<tr>
<td>Weissfluhjoch_out</td>
<td>Wind mast of SLF on rocky mountain peak</td>
</tr>
<tr>
<td>Eggishorn</td>
<td>Wind mast of SMI on local peak with rocky surroundings</td>
</tr>
<tr>
<td>Les Diablerets</td>
<td>Wind mast of SMI on mountain peak with rocky surroundings</td>
</tr>
<tr>
<td>Gornergrat</td>
<td>Wind mast of SMI on mountain peak with rocky surroundings</td>
</tr>
<tr>
<td>Jungfraujoehch</td>
<td>Research platform in high alpine environment (rock, snow, glacier)</td>
</tr>
</tbody>
</table>

### Table 3.4: Periods of observation.

<table>
<thead>
<tr>
<th></th>
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<th></th>
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<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>Locarno-Monti</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Payerne</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Davos See</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Davos</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Cimetta</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Männlichen</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>SLF Versuchsfeld</td>
<td></td>
<td></td>
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<td></td>
<td></td>
</tr>
<tr>
<td>Weissfluhjoch</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Weissfluhjoch_out</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Eggishorn</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Les Diablerets</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Gornergrat</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Jungfraujoehch</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

The nine mountain stations (stations above 1500 m a.s.l.) are all located on peaks or ridges, except the stations Davos and Versuchsfeld SLF. The peak- and ridge-stations have the advantage of a free horizon (APPENDIX A), but they are also exposed to the very harsh alpine weather conditions. The two other mountain stations and the two low altitude stations Payerne and Locarno-Monti have an acceptable horizon (maximum elevation always < 22°). The surroundings and the ground type of each station are described in Table 3.3.
Most of the stations have been operational for about four years (cf. Table 3.4) without heavy maintenance, except for the replacement of some ventilators. The oldest ASRB-stations have been operational since the end of 1994, the youngest (Davos-See) since the beginning of 1997. It is planned to move the station Davos from the roof of PMOD down to the station Davos-See, where the upward fluxes are measured. This will be done as soon as enough power is available there. At the moment the station Davos-See, which is only 200 m away from PMOD, is operational with the power from a solar panel at the bottom of the mast. The station Jungfraujoch was moved at the end of November 1997 from the old terrace to the new CHARM research platform and has been operated unshaded since then. The reason is that a shadow band would perturb other radiation measurements. The downward looking instruments of Weissfluhjoch out are located on the same arm as the upward looking instruments of Weissfluhjoch. They are defined as a separate station because they are only temporarily operational to investigate the difference in albedo between the ridge location of Weissfluhjoch and the valley location of SLF Versuchsfeld, which is only 500 m away from Weissfluhjoch.

3.2 Standard Instrumentation

The main goal of the ASRB network is to provide high quality radiation measurements at different altitudes despite the rough meteorological environment and remote locations, which do not allow regular maintenance. An important condition was therefore a fully automatic and reliable measurement system.

3.2.1 Pyranometer and Pyrgeometer

The standard instrumentation consists of a standard Kipp&Zonen CM21 pyranometer and a Eppley PIR pyrgeometer measuring shortwave global radiation and longwave atmospheric radiation, respectively. Both instrument types are one of the best available and their characteristics are explained in Chapter 2. The Eppley PIRs are modified, i.e. they are equipped with three dome thermistors, and calibrated at PMOD.

3.2.2 Ventilation and Shading

Both radiation instruments are mounted on an arm fixed about 2 m away from the main mast and about 4-8 m above the ground. Due to harsh meteorological weather conditions the development of a special ventilation system with preheated air to prevent hoar-frost and dew on the domes was an important issue. The metallic protection dome around the instrument body and the ventilation system provide a uniform air flow to the instruments and the domes. Ventilation air is sucked from below through a tube and preheated by a power resistor mounted below the ventilator. The axial ventilator on top of the instrument’s arm structure blows the air into the instrument chamber. Following the wall of the metallic protection shield, the air is then heated for the second time by an annular heating ring just before reaching the dome of the instrument (Figure 3.2). Thus the metallic shield and the ventilation system provide a uniform air flow to the instrument and its dome. The heating power is kept low to prevent perturbing temperature gradients in the instrument body.

As already mentioned, a pyrgeometer has to be shaded from the sun for accurate measurements but a shading disc on a tracker system is impossible at remote alpine stations. We therefore make use of a fixed vertical shadow band (5 cm width) in the southern direction and correct the longwave data for the rest of the day with an algorithm using the shaded data at noon (Chapter 4). The shadow band is white at the outer side (reduction of solar heating) and black at inner side (reduction of shortwave reflection).
3.2.3 Additional Measurements

ENET-stations are normally equipped with a VT100-thermometer to measure temperature; humidity is not measured. The stations Eggishorn, Männlichen and Diablerets (Gornergrat already had one) were deployed by SMI during fall 1996 with a Thyan - the high precision instrument of SMI to measure air temperature and humidity. All ASRB station have since then been equipped with a Thyan. Pressure data is available at seven ANETZ locations (Table 2.3). These standard meteorological elements are measured instantaneously every 10 minutes and the data are collected by SMI, but available for our use.

3.2.4 Data Acquisition

The measured signals are acquired by a Campbell CR10 datalogger, which is installed in a box at the bottom of the mast. A modem for the transmission of the data and a 12V power supply for the ventilation and heating system are also located in the same box. The availability of power and telephone lines was one of the reasons for which we deployed the ASRB station at already existing locations of ENET- and ANETZ network of SMI. All elements are measured every 5 seconds; every 2 minutes an average value is computed and saved with the time stamp of the end of the time interval on the datalogger (Until the end of 1995 only 5 minutes averages were stored). The data are transmitted daily via telephone lines from the stations to the central computer at PMOD/WRC in Davos.

3.3 Data-Collection

The data of each of the eleven stations are collected every four hours by the central computer at PMOD/WRC in Davos. A 2 minutes data array consist of the following parameters as listed in Table 3.5:
The ASRB-Network

Table 3.5: 2 minutes average parameters measured at each ASRB-station.

<table>
<thead>
<tr>
<th>Column #</th>
<th>Parameter</th>
<th>Notes / Unit</th>
<th>Unit</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>ID</td>
<td>Station number</td>
<td></td>
</tr>
<tr>
<td>2</td>
<td>Year</td>
<td>1994-??????</td>
<td></td>
</tr>
<tr>
<td>3</td>
<td>DoY</td>
<td>Day number (=Day of Year)</td>
<td></td>
</tr>
<tr>
<td>4</td>
<td>Time</td>
<td>0...2400</td>
<td>[UTC]</td>
</tr>
<tr>
<td>5</td>
<td>$U_{PYR}$</td>
<td>Voltage signal of the pyranometer</td>
<td>[mV]</td>
</tr>
<tr>
<td>6</td>
<td>$U_{PIR}$</td>
<td>Voltage signal of the pyrgeometer</td>
<td>[mV]</td>
</tr>
<tr>
<td>7</td>
<td>$T_{PYR}$</td>
<td>Body temperature of the pyranometer</td>
<td>[°C]</td>
</tr>
<tr>
<td>8</td>
<td>$T_{PIR}$</td>
<td>Body temperature of the pyrgeometer</td>
<td>[°C]</td>
</tr>
<tr>
<td>9</td>
<td>$T_N$</td>
<td>PIR dome temperature in direction north</td>
<td>[°C]</td>
</tr>
<tr>
<td>10</td>
<td>$T_{SE}$</td>
<td>PIR dome temperature in direction south-east</td>
<td>[°C]</td>
</tr>
<tr>
<td>11</td>
<td>$T_{SW}$</td>
<td>PIR dome temperature in direction south-west</td>
<td>[°C]</td>
</tr>
<tr>
<td>12</td>
<td>$I_H$</td>
<td>Power of the heating- and ventilation system</td>
<td>[A]</td>
</tr>
<tr>
<td>13</td>
<td>STD$_{PYR}$</td>
<td>Standard deviation of the PYR signal</td>
<td>[mV]</td>
</tr>
<tr>
<td>14</td>
<td>STD$_{PIR}$</td>
<td>Standard deviation of the PIR signal</td>
<td>[mV]</td>
</tr>
</tbody>
</table>

The power signal of the heating- and ventilation-system and the standard deviations of the radiation fluxes are monitored for data check reasons. Together with identification number, the year, the day-number and the time there are 14 parameters to be stored every 2 minutes. This makes about 55 Kb per day and station. The logger clocks are set to Universal Time Coordinated (UTC), which is one hour behind Central European Time (CET). At the ENET-stations the loggers are shared with the SLF. This implies that the time is CET and the logger storage is overwritten after about 30 hours due to lack of memory at these stations. The time of these files is transformed to UTC in the following quality control procedures.

3.4 Quality-Control

Quality control is an important step before analyzing any radiation data. The ASRB data are never corrected. However, data suspected to be erroneous are flagged and noted in a meta-file. A daily file of the past day is created for each station every morning by the central computer at PMOD. These daily files are saved as “level-1” files on our server. There the data are ready for the daily check, which is performed with the data analysis and programming tool Igor Pro from Wavemetrics™. Besides some automatic checks about the correct identification number, the date, the time and the array length, the data-control program mainly tests if the radiation quantities are within a certain predefined physically possible range, which are listed in Table 3.6.
TABLE 3.6: Physically possible intervals of the controlled parameters.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Lower Bound</th>
<th>Upper Bound</th>
</tr>
</thead>
<tbody>
<tr>
<td>SW↓ [W m⁻²]</td>
<td>-5</td>
<td>1500</td>
</tr>
<tr>
<td>LW↓ [W m⁻²]</td>
<td>50</td>
<td>500</td>
</tr>
<tr>
<td>SW↑ [W m⁻²]</td>
<td>-5</td>
<td>1500</td>
</tr>
<tr>
<td>LW↑ [W m⁻²]</td>
<td>50</td>
<td>600</td>
</tr>
<tr>
<td>STP↓ [W m⁻²]</td>
<td>-1</td>
<td>50</td>
</tr>
<tr>
<td>I₀ [A]</td>
<td>1</td>
<td>4</td>
</tr>
</tbody>
</table>

**Figure 3.3:** Daily plot of all ASRB-stations for checking the quality by eye. The unit of all radiation fluxes is W m⁻². The main part shows the shortwave radiation in blue (left scale), longwave radiation in red and the equivalent radiation temperature of the instrument in orange (right scale). The dip during solar noon is caused by the fixed shadow band. The spacing between the red and the orange curve is a good indicator for the cloud coverage. As is typical for a fall situation there seems to be fog in Payerne and clear-sky in the mountains. The upper part of a figure shows the short- and longwave 2 min. standard deviation in blue and red (right scale) and the current of the ventilation system in light-green (left scale). The lower part of the figures of the three stations measuring also the upward fluxes show the short- and longwave net fluxes in blue and red, the radiation balance in violet (left scale) and the albedo in dark-green (right scale).

A running median filter with a user specified threshold allows to find some “wild points”. In the course of all these tests no data are deleted, but the outliers are flagged (TABLE 3.7) in the generated “level-2”
file of the controlled data (Section 3.5). This is the first complete and checked daily file. It consists of the same columns as the level-1 file, but has four additional columns: two for the calculated short- and longwave radiation flux and two for the corresponding quality flags. The quality flag “1” is added in the next step, the level 3 file, where the data are corrected for different known influences (Chapt. 4).

Table 3.7: Meaning of the different flags in the level-2 file used for quality control.

<table>
<thead>
<tr>
<th>Flag</th>
<th>Meaning</th>
</tr>
</thead>
<tbody>
<tr>
<td>128</td>
<td>Missing</td>
</tr>
<tr>
<td>64</td>
<td>Physically Impossible</td>
</tr>
<tr>
<td>32</td>
<td>Ventilation Power too low</td>
</tr>
<tr>
<td>16</td>
<td>Wild Point</td>
</tr>
<tr>
<td>1</td>
<td>Corrected</td>
</tr>
<tr>
<td>0</td>
<td>Everything Fine</td>
</tr>
</tbody>
</table>

After the automatic quality control procedures the controlled data of every station are plotted on one sheet as the so-called daily-plot for a visual-check (Figure 3.3). All found outliers are already tagged with different colors. For an experienced observer it is possible to detect some additional irregularities or errors. These data can be tagged by selecting the involved interval or points with the mouse. Figure 3.4 and Figure 3.5 show screen dumps as examples of this step. These data also get a “physically impossible” flag.

Figure 3.4: Example of manually tagged data for the ASRB-station Davos. After a power disruption in the middle of the night and the following on-set, the instruments’ signal displayed a strange behavior, coming up only slowly to the normal level. The automatic quality control procedures flagged the missing data and the data beyond the given “physically possible” limits (see annotations on the right of the graph). The “bad” range can now manually be tagged with the mouse.

The time and duration of all gaps, outliers or otherwise flagged data are recorded in a monthly meta-file. The controlled and flagged data and the meta-file are finally stored per month and station in a database (approximately 2 Mb data per month and station).
More information about meteorological conditions to explain some strange irregularities can be obtained from the SLF software packet INFOBOX, which plots many parameters like the amount of new snow or wind speed in addition to the other standard meteorological measurements.

3.5 Data Preparation

The ASRB-data have to go through a five level data preparation, which is summarized in TABLE 3.8. The 2 minutes raw data consist of 14 parameters including the identification number (ID) of the station, the year, the day number (day of year = DoY) and the time. The level-2 data are filled (without gaps), controlled and flagged. The calculated radiation fluxes (SW, LW) and the quality control flags (FL\_PR, FL\_PR) are put in four additional columns. The detected irregularities are written to a monthly metafile. The level-3 data are the first corrected (CHAPTER 4) and therefore user-prepared monthly files.

The most important parameters of the station and the explanation to the data is given in the header of each file. The arrays are reduced by deleting some parameters, which are no longer used, but the flags are still included. We normally recommend to external users to delete all the data with flags ≥ 16. We plot this data as a monthly graphic (FIGURE 3.6), which is very practical for an overview or for identifying clear day periods.

**FIGURE 3.6:** Monthly plot of the level 3 data of the ASRB station Gornergrat for October 1999. The left scale in W m⁻² is for the shortwave global radiation (blue) and the right scale for the longwave downward radiation (red) and the appropriate instrument temperature (orange) in W m⁻².
Table 3.8: The different ASRB data level and their characteristics.

<table>
<thead>
<tr>
<th>Level</th>
<th>Freq.</th>
<th>Included parameters and elements</th>
<th>Length</th>
<th>Remarks</th>
</tr>
</thead>
<tbody>
<tr>
<td>001</td>
<td>2 min.</td>
<td>ID, Year, DoY, Time, U_{PYR}, U_{PIR}, T_{PYR}, T_{PIR}, 3 x T_D, I_H, 2 x StdDev</td>
<td>day</td>
<td>raw data</td>
</tr>
<tr>
<td>002</td>
<td>2 min.</td>
<td>ID, Year, DoY, Time, SW, LW, F_{PYR}, F_{PIR}, U_{PYR}, U_{PIR}, T_{PYR}, T_{PIR}, 3 x T_D, 2 x StdDev</td>
<td>day</td>
<td>gaps filled, controlled and flagged</td>
</tr>
<tr>
<td>003</td>
<td>2 min.</td>
<td>Year, DoY, Time, SW, LW, F_{PYR}, F_{PIR}, T_{PIR}, 2 x StdDev</td>
<td>month</td>
<td>+ manually corrected</td>
</tr>
<tr>
<td>004</td>
<td>10 min.</td>
<td>Date, Time, SW, LW, Temp, R_Hum</td>
<td>month</td>
<td>+ meteo elements</td>
</tr>
<tr>
<td>005</td>
<td>10 min.</td>
<td>Date, Time, SW, LW, Temp, R_Hum, CSI</td>
<td>month</td>
<td>+ clear sky index + Averaging of Temp and RHum</td>
</tr>
</tbody>
</table>

The step to the level-4 decreases the data volumes by averaging the 2 minutes data to 10 minutes values. Real date and time instead of year, DoY and time are used. The flag- and the standard deviation-columns are deleted, the air temperature (Temp) and the relative humidity (R_Hum) at the station are included instead. These two meteo-elements are measured instantaneously every 10 minutes. To use them correctly together with the averaged 10 minutes radiation fluxes, we re-calculate them as follows: for the level-5 data we hence calculate the 10 minutes meteo-elements as the average of the previous and the actual value. Furthermore the calculated clear-sky index (CSI) (Chap. 5) is included (175 Kb per month and station).

3.6 Maintenance

Regular maintenance of the instruments over long periods is essential for detecting trends and will be a major challenge. We decided to control the long-term stability of the ASRB radiation instruments by comparing them with a “travelling standard”. This is a small mobile data acquisition system and radiation measurement platform with a carefully calibrated and maintained pyrgeometer and pyranometer. The problems and possible errors are much smaller if we leave the same instruments all the time at the stations and compare them to a travelling standard, instead of exchanging them. The stability of the travelling standard instruments can easily be checked with our calibration facilities at PMOD.

With good design, an absolute accuracy of less than one per cent is possible, but is difficult to improve substantially upon this figure. The more accurate the measurements, the easier it will be to interpret them. Within the scope of a recent international Round Robin pyrgeometer calibration experiment, five PIR pyrgeometers were calibrated at the beginning and the end of the experiment (four years later) by the same laboratory (Philipona et al. 1998). The calibration factors obtained were the same within a 1% limit, although the instruments had traveled to many different and distant locations. This provides a strong indication for the stability of PIR pyrgeometers.

Figure 3.7 shows the results of a visit with the Travelling Standard at the ASRB station Männlichen. The comparison started just after noon on 9 September 1997 and ended at noon the day after. During the night the deviation of the longwave downward radiation was in the range of ±2 W m⁻². With sunshine the difference gets larger since neither of the instruments were shaded. The deviation of the shortwave global radiation has a diurnal behavior, which is a sign of a slight tilt of a one of the pyranometers. The difference between the two pyranometers is also < 1%. The Travelling Standard was occasionally shaded around noon for experimental reasons. This caused the deviations around noon. As a whole, the comparison demonstrates that the instruments at this remote station still measure very accurately, despite the fact that they were exposed for 3 years without any maintenance.
Every station has been visited at least once during the last four years with the Travelling Standard (Figure 7.13). These checks and some data quality control procedures showed that the high quality equipment of the stations can fulfill the high demands and guarantee the desired accuracy. Dome pollution is particularly small at the mountain stations due to many storms, clean air and a filter within the ventilation system. Frequent control observations (visual by eye and by web-cam) at the manned stations Weissfluhjoch and the old Jungfraujoch location demonstrated that our ventilation and heating system effectively prevents the snow accumulation on the domes. However, the new Jungfraujoch location at the research terrasse is very exposed causing covered domes after blizzards.
4 Data Correction and Evaluation

Due to the complexity of the radiation measurements and the climatological demand for very accurate data, we have to take into account all the internal and external perturbations, which influence the measurements.

4.1 Longwave Downward Radiation

4.1.1 Direct Sun Influence

Introduction
As already mentioned, it is not possible to permanently shade pyrgeometers at the ASRB stations for technical reasons. The thereby increased dome heating is not really a problem because the modified ASRB-PIRs with the three dome thermistors correctly measure the dome temperature, which allows to compensate the dome effect. However, an unshaded pyrgeometer also measures the part of the direct solar radiation spectrum above the cut-on of the pyrgeometer dome (Figure 2.1). This cut-on, usually located between 3 and 4 μm, is different from instrument to instrument (SECTION 2.4.4).

TABLE 4.1: MODTRAN calculated amount of direct solar irradiance for wavelength above the cut-on of pyrgeometer domes for the two ASRB-stations Davos and Jungfraujoch at the beginning of August (day# 218). High solar elevation and low water vapor content (large atmospheric window) are the two major components responsible for high values of solar radiation unintentionally measured with a pyrgeometer. The largest values can hence be calculated at high stations during summer time.

<table>
<thead>
<tr>
<th>Cut-on of the dome</th>
<th>Davos, 1610 m a.s.l. Mid.Lat.Summer Atmosphere day# 218, zenith angle 25°</th>
<th>Jungfraujoch 3580 m a.s.l. Sub.Arc. Winter Atmosphere day# 218, zenith angle 25°</th>
</tr>
</thead>
<tbody>
<tr>
<td>&gt; 2.5 μm</td>
<td>12.7 W m⁻²</td>
<td>23.4 W m⁻²</td>
</tr>
<tr>
<td>&gt; 3.0 μm</td>
<td>11.8 W m⁻²</td>
<td>18.4 W m⁻²</td>
</tr>
<tr>
<td>&gt; 3.5 μm</td>
<td>7.9 W m⁻²</td>
<td>10.9 W m⁻²</td>
</tr>
<tr>
<td>&gt; 4.0 μm</td>
<td>3.3 W m⁻²</td>
<td>5.6 W m⁻²</td>
</tr>
<tr>
<td>&gt; 4.5 μm</td>
<td>2.2 W m⁻²</td>
<td>4.3 W m⁻²</td>
</tr>
<tr>
<td>&gt; 5.0 μm</td>
<td>1.1 W m⁻²</td>
<td>2.6 W m⁻²</td>
</tr>
<tr>
<td>&gt; 6.0 μm</td>
<td>0.9 W m⁻²</td>
<td>1.5 W m⁻²</td>
</tr>
</tbody>
</table>

The amount of this direct solar radiation unintentionally transmitted to the sensor of the PIR depends not only on the blocking edge of the dome. The solar transmittance spectrum itself varies with the atmospheric water vapor content and the solar elevation, which both depend on season and location. Measurements (FIGURE 4.1 and FIGURE 4.6) and calculations (TABLE 4.1) demonstrate that this amount of direct solar radiation can rise up to more than 20 W m⁻². The impact of the direct solar radiation is therefore not at all negligible and has to be corrected.
FIGURE 4.1: Example of the longwave shadow dip at solar noon. The measured influence of the direct solar radiation is about 15 W m⁻² for this June day at the ASRB station Maennlichen. This quantity is not at all negligible and has to be corrected.

For the corrections we make use of the south oriented, fixed vertical shadow band at the ASRB-stations. At solar noon it casts a uniform shadow over the domes for 10 minutes, independent of the declination. On sunny days the amount of the direct solar radiation, which has to be subtracted from the measurement, can be determined during the shaded period.

**Old Correction Method**

Philipona at al. (1995) introduced the dome temperature difference term

\[
\Delta T_{S-N} = (T_{SE} - T_N) + (T_{SW} - T_N),
\]

where the subscripts \(N, SE, SW\) stand for the direction of the three thermistors in the dome of modified pyrgeometers. They showed that \(\Delta T_{S-N}\) correlates with the global radiation (Figure 4.2). Possible offsets between the three thermistors have to be canceled for this method by matching the three thermistors to each other using night time measurements.

They defined a correction factor \(f\), which can be multiplied with \(\Delta T_{S-N}\) and used as a measure for the amount of the direct solar "longwave" radiation above the cut-on of the dome. Direct solar irradiance has a local and seasonal dependence (humidity and solar elevation). This factor \(f\) is therefore not a constant, but inherent to the instrument and to the solar angle. The actual value of \(f\) can be determined during a clear day from the amount of the direct solar "longwave" radiation measured as a difference between the shaded and the adjacent unshaded values of the pyrgeometer around solar noon. As a result they defined the correction term to be subtracted as

\[
f \cdot \Delta T_{S-N}
\]
Attempts to use this method operationally within the ASRB-network failed for two reasons. First, average values for $\Delta T_{S-N}$ are approximately 0.3K during noon and the accuracy of the thermistors is only approximately 0.1K. The consequences of such small temperature differences values may lead to wrong results when subtracting the single terms. Also, due to slightly different heat conduction between thermistors and domes, the individual dome temperatures may not be balanced. Furthermore, in summer (especially in high latitudes) the diurnal $\Delta T_{S-N}$ does not follow the solar irradiance due to the fact that the sunrise is in NNE and the sunset in NNW, which both warms the N-thermistors during these times more than the SE- or SW-thermistor. All these reasons result in wrong and asymmetric diurnal $\Delta T_{S-N}$ curve (FIGURE 4.3), which can not be used for the correction of the direct solar influence on the pyrgeometer.

**FIGURE 4.2:** Diurnal evolution of the dome temperature difference between south and north $\Delta T_{S-N}$ compared with the global radiation at the ASRB-station Gornergrat.

**FIGURE 4.3:** Same as FIGURE 4.2, but for the ASRB-station Locarno-Monti and summer. The $\Delta T_{S-N}$ term correlates badly with the measured global radiation (SW). The reason for the asymmetric $\Delta T_{S-N}$ curve is the different heat conduction of the thermistors and the high sun elevation influencing also the neighbouring thermistors.
New Correction Method

The best element to correct for the influence of direct solar radiation on pyrgeometer measurements would be the measurement of direct solar irradiance itself. Unfortunately this element is not measured at the ASRB station, nor is it measured at most radiation stations worldwide. Global solar radiation is the widely measured radiation parameter and in many cases it is comparable to the direct solar radiation on a horizontal surface, due to the small amount of diffuse radiation during clear days. Unfortunately, this is not true for mountainous regions like the Alps, since snow and low clouds produce high diffuse solar radiation.

![Graph showing correction of direct solar radiation](image)

**Figure 4.4:** Correction of the direct solar radiation unintentionally measured by the pyrgeometer at the ASRB-station Davos with the help of the calculated direct solar irradiance (violet). The corrected longwave downward flux (dark-blue) is compared to a permanently shaded pyrgeometer (dark-green). Although the cloud cover and thus the direct solar irradiance changes heavily in the afternoon the difference (light-green) between the corrected ASRB pyrgeometer and the shaded pyrgeometer is always within ±2 Wm⁻².

After investigating many different possibilities we decided to approximate the direct solar irradiance with the help of the global radiation measured with the pyranometer. During solar noon the pyranometer is also shaded by the fixed shadow-band. Hence, the pyranometer measures only diffuse solar radiation during these 10 shaded minutes. Direct solar irradiance at noon can therefore be calculated as the difference between the shaded and the adjacent unshaded measurements of the pyranometer during solar noon. The variation of the diffuse radiation during a day is normally small in comparison to the global radiation. The diurnal course of diffuse radiation was therefore estimated with a cubic spline interpolation between the noon value and the values before and after sunrise/sunset. The diurnal course of direct solar radiation could then be determined calculating the difference between the measured global and approximated diffuse radiation. This approximated direct radiation $SW_{dir}$ is divided by 1000...
to be able to use it in connection with a reasonable factor $f$. As a result, the new correction term can be written as

$$f \cdot (SW_{dir} / 1000)$$

Figure 4.4 shows the result of the method for a typical summer day at the ASRB station Davos. The measured global radiation indicates a clear morning followed by an afternoon dominated by a changing convective cloud cover. The shadow dip at solar noon (11:30 UTC) displays a scaled diffuse radiation of about 0.15 (i.e., 150 W m$^{-2}$). This value is used to interpolate the diffuse curve. The resulting approximated direct solar irradiance can now be used to correct for the amount of solar radiation above the cut-on of the pyrgeometer dome.

At the station Davos, we are able to check the results of the proposed method with a permanently shaded pyrgeometer. The ASRB measured longwave downward radiation (Figure 4.4) shows a clear dip at solar noon, indicating an error of approximately 12 W m$^{-2}$ due to the direct sun. An $f$-factor of 15 had to be used for this day to find the correct longwave downward curve, which just matches the shadow dip. The small difference between the corrected ASRB pyrgeometer and the shaded pyrgeometer curve demonstrates that the approximation of $SW_{dir}$ is quite good (within ±2 W m$^{-2}$).

The method even works satisfactorily if the sky is overcast at noon and clear in the afternoon, because the determined diffuse value at solar noon is equal to the measured global radiation. The largest problems with this method have been found for the few cases where the noon is clear, but a fresh snow layer on the surface and a very thin, low fog layer produces very large diffuse radiation before or/and after noon.

The correction just before and after the shading is a minor problem. The fact that the pyranometer sensor is larger than the pyrgeometer sensor results in a wider shadow dip in the shortwave (Figure 4.1). The time of non-correction of the longwave downward radiation is therefore a little bit too long. This problem has been taken care of by interpolating the values around solar noon.

**Finding the Proper Correction Factor $f$**

The current $f$-factor can easily be calculated during a clear solar noon with the following equation:

$$f = (LW_{noon \text{max}} - LW_{noon \text{min}}) / SW_{dir, noon \text{max}}$$

Whereas $(LW_{noon \text{max}} - LW_{noon \text{min}})$ is the difference between the shaded $LW_{dir}$ and the adjacent unshaded $LW_{dir}$, $SW_{dir, noon \text{max}}$ is the approximated unshaded direct solar irradiance around noon. The correction factor $f$ is thus a stretch factor to be multiplied with $SW_{dir}$ to find the amount of unintentionally measured solar radiation (Equation 4.3).

To operationally use the $f$-factor, we have to know it for every day of a year, independent of clear or cloudy conditions. The seasonal dependence (solar angle) can be omitted, because it is already considered by the direct solar irradiance itself. The new $f$-factor is no longer strongly dependent on the total water vapor quantity in the atmosphere, since the water vapor concentration in the atmosphere not only influences the longwave absorption but also the scattering of the direct solar irradiance. The $f$-factor is therefore more or less constant throughout the year (Figure 4.5). The amount of the direct solar irradiance to be corrected, in contrast, has a strong seasonal cycle (cf. Figure 4.6) due to the changing direct solar irradiance.

The value of the $f$-factor can thus easily be found after calculating the $f$-factor for approximately a dozen clear noons with different atmospheric conditions. Clear noons can be identified numerically by a strong shadow dip and small standard deviations in the solar radiation before and after the dip. The more clear noons, the more different atmospheric conditions, and the better the average of the calcu-
lated $f$-factors. We determined the $f$-factor for every ASRB-station by averaging all found values throughout a year (Table 4.3).

**Figure 4.5:** Calculated $f$-factors for all clear noons during 1997 at the ASRB station Locarno-Monti. The mean $f$-factor of this time-series is taken as the characteristic $f$-factor for the pyrgeometer at this location.

**Figure 4.6:** Annual course of the amount of direct solar irradiance unintentionally measured (and thus to be corrected) with the pyrgeometer at the ASRB-station Locarno-Monti at clear noons during 1997. The highest values occur in summer due to high solar elevations.

Investigations showed that the day to day variations of the $f$-factor due to changing water vapor in the atmosphere are relatively small and may be ignored. Nevertheless, we consider this variation by comparing the current clear day $f$-factor with the averaged $f$-factor. After checking the quality of the found current $f$-factor, the final daily $f$-factor is determined by calculating the average between the current and averaged annual $f$-factor.
The explained determination of the $f$-factor is a major advantage of this new method in comparison to the old one. There, the $f$-factor was really seasonal dependent and had therefore to be determined with a sine-fit through all the clear noon values throughout the year.

### 4.1.2 Influence of the Shadow Band

The shadow band not only influences the pyrgeometer measurements during noon. Together with the main mast, the shadow band blocks some sky radiation which has to be corrected for. Philipona et al. (1995) introduced a correction factor $g$, by which the pyrgeometer voltage signal $U_{PIR}$ has to be multiplied to account for this effect (assuming isotropic sky radiance).

\[
g \cdot U_{PIR}.
\]  

The factor $g$ is dependent on the percentage of the hemisphere, which is covered by the shadow band and the mast. It can be determined by a geometrical calculation and is thus a constant - assuming that the shadow band (and mast) temperature is the same as the ambient temperature. The shadow band is always warmer than the sky, causing a too high signal. This too high measurement needs to be corrected by subtracting the influence of the shadow band (and the mast). However, the $U_{PIR}$ of a sky facing pyrgeometer is always negative, since the pyrgeometer thermopile loses heat. The correction term $g \cdot U_{PIR}$ therefore has to be added to the pyrgeometer formula. In normal circumstances the effect of shadow band and mast is maximum 2 W m$^{-2}$.

The pyrgeometer signal $U_{PIR}$ of a downward looking pyrgeometer (no shadow band) can be slightly positive, if the observed surface gets warmer than the surrounding air. The effect of the more or less equally warmed mast can be ignored in this situations.

### 4.1.3 Pyrgeometer Formula With Correction Terms

The pyrgeometer formula (EQUATION 2.6) with two correction terms can now be written as:

\[
LW_\downarrow = \frac{U_{PIR}}{C_{PIR}} \left( 1 + k_1 \sigma T_B^4 \right) + k_2 \sigma T_B^4 - k_3 \sigma (T_D^4 - T_B^4) - fSW_{DIR} + gU_{PIR}.
\]  

Both correction terms are added in accordance with the proceeding discussions. The effect and accuracy of the two corrections are summarized in the following TABLE 4.2.

**Table 4.2: Effect and accuracy of the two corrections concerning the pyrgeometer measurement.**

<table>
<thead>
<tr>
<th>Correction</th>
<th>Range</th>
<th>Accuracy</th>
</tr>
</thead>
<tbody>
<tr>
<td>Direct solar irradiance above the cut-on of a PIR dome [W m$^{-2}$]</td>
<td>5 - 20</td>
<td>1 - 2</td>
</tr>
<tr>
<td>Shielding effect of the shadow band [W m$^{-2}$]</td>
<td>max. 2</td>
<td>0.2</td>
</tr>
</tbody>
</table>

The correction factors used within the ASRB-network are compiled in TABLE 4.3. The $f$-factor values give an evidence for the location of the pyrgeometer dome cut-on. High values indicate low cut-on and vice-versa. The $g$-factors are very different, although the shadow band is the same at all ASRB-stations. The reason is the different heights and thickness of the masts or other constructions nearby. The $g$-factors for the downward looking pyrgeometers are zero or small, because the temperature difference between the mast and the ground is close to zero.
4.2 Shortwave Global Radiation

4.2.1 Noon Shadow

The shadow band is necessary for the longwave measurement. The shading of the pyranometer during solar noon additionally provides useful information about diffuse solar radiation. The relationship between the diffuse and direct solar radiation, for example, is a fair indicator for atmospheric turbidity. This noon shadow dip has to be corrected for radiation balance investigations. Proper clear noons (low standard deviations before and after) are interpolated with a polynomial fit. Noons with a weak shadow dip and large standard deviations before and after are interpolated with a linear fit.

4.2.2 Zero-Offset

The physics of the zero-offset is explained in Section 2.3.2. Due to an effective heating and ventilation system at the ASRB-stations, zero-offsets usually do not exceed -2 W m\(^{-2}\) (Figure 4.7). Values larger than -5 W m\(^{-2}\) are seldom observed and only in connection with improper functioning of the ventilation system. The largest offsets have been measured at the highest and very exposed station Jungfraujoch during clear nights after a storm. After such conditions the ventilation system can be reduced and the heated air blocked between the instrument and the protection dome by snow and ice.

Zero-offsets larger than -5 W m\(^{-2}\) are therefore flagged in the level-3 files. Zero-offsets smaller than -5 W m\(^{-2}\) are simply shifted to zero during the night and no further adjustments made for the data obtained during the day. The reason for applying no zero-offset correction for the daytime data is that

---

**Table 4.3**: Correction factors of all ASRB-stations used for the operational data correction within the data reduction algorithms. Upward fluxes do not have to be corrected for the direct sun influence, the f-factors are therefore zero at these stations. The g-factors (for upward fluxes) in brackets are only valid when the voltage signal \(U_{PR}\) of the pyrgeometer is negative, i.e. the surface is colder than the air.

<table>
<thead>
<tr>
<th>ASRB-station</th>
<th>Altitude</th>
<th>f-factor</th>
<th>g-factor</th>
</tr>
</thead>
<tbody>
<tr>
<td>Locarno-Monti</td>
<td>370 m</td>
<td>12.7</td>
<td>5</td>
</tr>
<tr>
<td>Payerne</td>
<td>490 m</td>
<td>11.7</td>
<td>4</td>
</tr>
<tr>
<td>Davos See</td>
<td>1550 m</td>
<td>-</td>
<td>(5)</td>
</tr>
<tr>
<td>Davos</td>
<td>1610 m</td>
<td>13.5</td>
<td>4</td>
</tr>
<tr>
<td>Cimetta</td>
<td>1670 m</td>
<td>13.3</td>
<td>4</td>
</tr>
<tr>
<td>Männlichen</td>
<td>2230 m</td>
<td>15.5</td>
<td>8</td>
</tr>
<tr>
<td>SLF-Versuchsfeld_in</td>
<td>2540 m</td>
<td>12.8</td>
<td>4</td>
</tr>
<tr>
<td>SLF-Versuchsfeld_out</td>
<td>2540 m</td>
<td>-</td>
<td>(2)</td>
</tr>
<tr>
<td>Weissfluhjoch</td>
<td>2690 m</td>
<td>12.0</td>
<td>5</td>
</tr>
<tr>
<td>Weissfluhjoch_out</td>
<td>2690 m</td>
<td>-</td>
<td>(3)</td>
</tr>
<tr>
<td>Eggishorn</td>
<td>2895 m</td>
<td>14.5</td>
<td>6</td>
</tr>
<tr>
<td>Les Diablerets</td>
<td>2965 m</td>
<td>14.1</td>
<td>6</td>
</tr>
<tr>
<td>Gornergrat</td>
<td>3110 m</td>
<td>11.8</td>
<td>6</td>
</tr>
<tr>
<td>Jungfraujoch</td>
<td>3580 m</td>
<td>14.9</td>
<td>3</td>
</tr>
</tbody>
</table>
the depression for the daytime observation hours is very uncertain (SECTION 2.3.2) and the nighttime offset falls within the standard deviation of the calibration coefficient.

Investigations demonstrated that the shortwave global radiation is plus/minus zero when the sun-elevation is approximately -5°. All shortwave values with the solar elevation less than -5° are set to zero in the level-3 files.

![Graph](image)

**FIGURE 4.7**: Example for the very low zero-offset of the ASRB-pyranometers. The measured longwave downward radiation is about 160 W m⁻² during this cold night, which causes a fair cooling of the pyranometer dome. Nevertheless, the observed zero-offset never exceeds -2 W m⁻² due to the very effective heating and ventilation system.

### 4.3 Comparison With Independent Data

At the Payerne station we have the possibility of comparing the ASRB-data with BSRN-data, because Payerne is also a BSRN-station. The quality requirements for a BSRN station are very high and the maintenance of the radiation instrument is therefore excellent. The short- and longwave fluxes are measured at the same location with the same type of instruments. The data frequency (1 min.) and the ventilation system is only slightly different. The conditions for a direct comparison are therefore ideal.

The BSRN pyrgeometer is also modified (with three dome thermistors), but permanently shaded with a rotating shadow disc. The BSRN pyranometer operates unshaded. The correction of the direct solar radiation on the pyrgeometer dome and the noon shadow correction on the pyranometer can thus be omitted for the BSRN data.

**FIGURE 4.8** and **FIGURE 4.9** show that differences of the monthly mean values between the ASRB- and BSRN-data are within 0.3%. The exceptions are two months during which a not negligible amount of data was missing. This result demonstrates that it is possible to measure the longwave downward radiation very accurately with a fixed shadow band and that the correction algorithms for the ASRB-data work very well for the short- and longwave radiation.
FIGURE 4.8: Differences between monthly mean shortwave global radiation ASRB- and BSRN-data at Payerne.

FIGURE 4.9: Differences between monthly mean longwave downward radiation ASRB- and BSRN-data at Payerne.
Detecting Clear-Sky Situations

5.1 What is "Clear-Sky" and why do we Need it?

The term "clear-sky" is used in the context of this work to distinguish a sky without any clouds and opacity from a sky with hydrometeors like clouds. The term "cloudy sky" is used, if the hemisphere is covered with some clouds or totally overcast. The term "all-sky" is used for the total scene, i.e., all observed conditions.

Many investigations on radiation in the atmosphere require for the distinction between clear and cloudy sky situations. Most atmospheric models are first tested for clear-sky conditions, because the atmosphere itself is complex enough without clouds. Within the ASRB project most investigations where first accomplished with clear-sky data, because correlations and dependencies between the different parameters can more easily be understood. Since we are also interested in the cloud cover and its characteristics we do not at all exclude cloudy situations. An important application is the calculation of the cloud forcing (Chapt. 7).

Furthermore, the long-term changes in cloud amount and the resulting effect on the radiative energy balance is one of the largest uncertainties in global climate change research. Only satellites provide the global coverage needed to study the cloud amount on a global scale. However, as with other satellite derived variables, the accuracy of satellite cloud cover fraction retrievals must be validated with surface measurements.

5.2 Possibilities and Problems of Detecting Clear-Sky

Clouds can be thick or very thin and their base can be high or low. The cloud coverage can be homogeneous or irregularly broken, in one layer or distributed over several layers. The oldest approach to analyze cloud cover has been largely dependent on surface observer reports, which include temporal and spatial mismatches and observer subjectivity, adding uncertainty to the observation. Some automatic instruments have been developed, like a Hemispheric Sky Imager (Long and DeLuisi 1998), which allow to continuously monitor the cloud amount. Only very few radiation stations are equipped with such kinds of expensive and delicate instruments. A successful method should therefore only use standard measured meteorological or radiation elements. Recent studies estimate the fractional sky cover from broadband shortwave radiometer measurements (Long et al. 1999). These authors determined a relationship between sky cover amount and diffuse cloud effect, which they define as the measured diffuse shortwave irradiance minus the clear-sky diffuse shortwave irradiance. All approaches using shortwave data (thus also images in the visible wavelength) have the big disadvantage that an estimation of the cloud cover can only be accomplished during daytime.

In order to distinguish between clear-sky and cloudy situations during the entire day, a method based on longwave radiation measurements is more appropriate. The principle of the approach is a simple comparison between the measured longwave downward radiation and the calculated theoretical clear-sky longwave downward radiation at the same moment (Ambrosetti 1991). The method is thus more or less the reversed approach of all the attempts to calculate the $LW_{\downarrow}$ out of meteorological surface measurements and synoptic sky observations.
5.3 The Clear-Sky Index

As already explained in Figure 3.3 the cloud cover of a sky can be visually approximated by comparing the air temperature (computed in W m$^{-2}$) with the measured LW$\downarrow$. That is more or less similar to the above mentioned method to look at the relationship between the actually measured longwave downward radiation and a theoretical clear-sky longwave downward radiation. To use this approach on an operational basis, the diurnal and seasonal variations of LW$\downarrow$ have to be normalized. This can be achieved by comparing the apparent emissivity (Unsworth and Monteith 1975) instead of the LW$\downarrow$. Considering the atmosphere as a grey body, the apparent emissivity $\varepsilon_A$ can be estimated with the help of the Stefan-Boltzmann law:

$$\varepsilon_A = \frac{LW\downarrow}{\sigma T_a^4},$$  \hspace{1cm} (5.1)

where $\sigma$ is the Stefan-Boltzmann constant and $T_a$ [K] the air temperature. LW$\downarrow$ is thus normalized with the black body radiation flux of the air temperature.

The apparent emissivity for a clear-sky, in contrast, can be calculated very accurately, if the vertical temperature and water vapor profiles are known. In practice such profiles are usually not available and the available data are limited to standard meteorological elements measured at the surface. There are many empirical equations in the literature (König-Langlo and Augstein 1994, Auelinet 1994, Ramsey et al. 1981, Idso 1981), which calculate the apparent emissivity with the help of temperature and/or humidity. After numerous investigations we decided to use the Brutsaert formula (Brutsaert 1975):

$$\varepsilon_{AC} = k(e_a/T_a)^{1/7},$$  \hspace{1cm} (5.2)

where $\varepsilon_{AC}$ is the apparent clear-sky emissivity, $k$ is a location dependent coefficient, which can be found with the help of some manually selected clear-sky data, $e_a$ is the water vapor pressure [Pa] and $T_a$ [K] the air temperature at screen level height. Brutsaert derived this equation analytically by integrating Schwarzschild’s transfer equation for simple atmospheric profiles of temperature and water vapor pressure, and prescribing a simple dependence of the emissivity on the water vapor amount. A limitation of this expression is that the effect of greenhouse gases other than water vapor is neglected (Konzelmann 1994). To take this effect into account the formula is somewhat expanded by a variable $\varepsilon_{AD}$, which is the altitude dependent clear-sky emittance of a completely dry atmosphere as calculated with MODTRAN. Absorption bands of water vapor and other gases overlap. Therefore, $\varepsilon_{AC}$ should increase more slowly with $e_a$ than if greenhouse gases other than water vapor are neglected. Hence, the exponent is expected to be smaller or equal to 1/7. Indeed, an optimal fit to data of different stations was obtained by setting the exponent to 1/8. After this modifications our formula for calculating the theoretical clear-sky apparent emissivity $\varepsilon_{AC}$ can be written as:

$$\varepsilon_{AC} = \varepsilon_{AD} + k(e_a/T_a)^{1/8}. \hspace{1cm} (5.3)$$

More explanation to $\varepsilon_{AD}$ and $k$ can be found in the next section. Since we only want to know whether the sky is clear or not at a given moment, we defined a Clear-Sky Index (CSI), which is based on the relation of $\varepsilon_A$ to $\varepsilon_{AC}$:

$$\text{CSI} = \frac{\varepsilon_A}{\varepsilon_{AC}} \hspace{1cm} (5.4)$$

The “measured” apparent emissivity is thus normalized with the theoretically possible clear-sky apparent emissivity at same conditions. Therefore, if we find CSI ≤1 the sky should be clear and otherwise, if CSI >1 there should be some clouds around (cf. Table 5.1). The CSI is calculated every 10 minutes, because the air temperature $T_a$ is only measured with this frequency.
TABLE 5.1: Definition of the Clear-Sky Index (CSI) based on Equation 5.4

<table>
<thead>
<tr>
<th>CSI</th>
<th>Sky-Cover</th>
</tr>
</thead>
<tbody>
<tr>
<td>≤1</td>
<td>clear-sky, no clouds</td>
</tr>
<tr>
<td>&gt;1</td>
<td>cloudy sky, overcast</td>
</tr>
</tbody>
</table>

5.4 Finding the Proper Local Coefficients

5.4.1 The Altitude Dependent $e_{AD}$ Factor

$e_{AD}$ is the clear-sky emittance of a completely dry atmosphere as explained in Section 5.3. Calculations with MODTRAN showed a slight altitude dependence of $e_{AD}$, since the partial pressure of all gases is decreases with increasing altitude. We did the analysis with the MODTRAN integrated mid-latitude summer and winter atmosphere model. The differences between the two atmospheric models were quite small, due to the fact that the impact of the seasonal differences of the greenhouse gases other than water vapor are fairly small. We therefore used the following (Table 5.2) averaged values for the different altitudes.

**Table 5.2: MODTRAN calculated clear-sky emittance of a completely dry atmosphere ($CO_2=360$ ppm). This emittance of all greenhouse gases other than water vapor is slightly altitude dependent.**

<table>
<thead>
<tr>
<th>ASRB-station</th>
<th>Altitude</th>
<th>$e_{AD}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Locarno-Monti</td>
<td>370 m</td>
<td>0.23</td>
</tr>
<tr>
<td>Payerne</td>
<td>490 m</td>
<td>0.23</td>
</tr>
<tr>
<td>Davos</td>
<td>1610 m</td>
<td>0.22</td>
</tr>
<tr>
<td>Cimetta</td>
<td>1670 m</td>
<td>0.22</td>
</tr>
<tr>
<td>Männlichen</td>
<td>2230 m</td>
<td>0.22</td>
</tr>
<tr>
<td>SLF-Versuchsfeld</td>
<td>2540 m</td>
<td>0.21</td>
</tr>
<tr>
<td>Weissfluhjoch</td>
<td>2690 m</td>
<td>0.21</td>
</tr>
<tr>
<td>Eggishorn</td>
<td>2895 m</td>
<td>0.21</td>
</tr>
<tr>
<td>Les Diablerets</td>
<td>2965 m</td>
<td>0.21</td>
</tr>
<tr>
<td>Gornergrat</td>
<td>3110 m</td>
<td>0.21</td>
</tr>
<tr>
<td>Jungfraujoch</td>
<td>3580 m</td>
<td>0.20</td>
</tr>
</tbody>
</table>

5.4.2 The Location Dependent $k$-Coefficient

Having established the type of expression for the apparent clear-sky emittance (Equation 5.3), an optimal fit to the available data for every station was made. The problem here is that we would like to find the clear-sky situation with our method, but we first have to find the station coefficient $k$ with the help of clear-sky data.

Stations with Cloud Observations

For the ANETZ-stations we used the synoptic sky observations, which are usually made three or four times per day. The observed cloud amounts, which are instantaneous values, were associated with
hourly means (centered around the cloud observation at x:30) of longwave downward radiation, air
temperature and water vapor pressure. For the analysis of the clear-sky emittance, samples were
selected whenever the cloud amount was less than or equal to two eighths.

At the ASRB-station Locarno-Monti, for example, the selected clear sky data-set of the year 1997 con¬
tained 761 samples, with which $\varepsilon_{AC} = \frac{LW^{\prime}}{\sigma T_a^4}$ was computed. The resulting values are plotted
against the ratio of $e_a/T_a$ in EQUATION 5.3. A minimum standard deviation of the residuals was found
in this case for $k = 0.440$ and the corresponding curve is shown in FIGURE 5.1. Apparently
EQUATION 5.3 is fairly well confirmed by the data, which justifies a clear-sky emittance that is depen¬
dent of $e_a/T_a$.

All points in FIGURE 5.1 are measured during clear-sky situations. The determined EQUATION 5.3 rep¬
resents only an averaged parameterization for the clear-sky emittance at this station. Our method to auto¬
matically detect clear-sky requires a curve on a higher level, which follows the maximum clear-sky
emittance observed at this station. After many investigations with data from all stations, we defined
this curve by the upper two sigma confidence band (dashed curve in FIGURE 5.1). We applied this
method for the three years 1995-1998 and found very similar $k$-coefficients for every year.

![FIGURE 5.1: Clear-sky emittance as calculated from measured 10 minutes values of longwave
downward radiation as a function of the ratio of screen-level water vapor pressure and
temperature. The curve represents the parametrization of the clear-sky emittance optimized by
means of these data.]

Stations without cloud observations

Often cloud observation data is not available. For about half of the ASRB-stations we had to find the
$k$-coefficient with another method. One possibility is to use the information of the fixed shadow band
to find clear sky noons as explained in SECTION 4.1.1. An other possibility is to manually select
clear-sky days based on a daily plot. Although the data amount is heavily reduced with both
approaches, they both work fine for the most stations. One exception is Payerne, where we very often
observe fog in the winter time, leading to very few clear-sky situations.

A third possibility is to plot all the data and delete (with a graphical user interface) all the cloudy-sky
conditions manually. But which points are the "cloudy" ones? As shown in FIGURE 5.2 the points are
concentrated at the lower and upper bound of the apparent sky emissivity for each ratio of $e_a/T_a$. The
upper bound concentration is caused by overcast conditions. The lower bound concentration is the one
we are interested in. Low apparent sky emissivity indicates clear-sky situations, to which we would like to fit Equation 5.3.

![Graph showing emissivity vs. water vapor pressure/temperature](image)

**FIGURE 5.2:** Same plot as Figure 5.1, but for the ASRB-station Gornergrat and all-sky conditions, because there is no synoptic sky observation data available to exclude the cloudy-sky situations. The density of the points is concentrated in the lower (0.5-0.6 emissivity) and upper (0.9-1 emissivity) part of the diagram, indicating clear (low emissivity) and overcast (high emissivity) conditions. After manually deleting the cloudy points the clear sky data can be fitted according to Equation 5.3.

**TABLE 5.3:** $k$-coefficients for the different ASRB. At stations with synoptic sky observations the $k$-coefficient was determined with the help of the observed clear sky situations. At all other stations the method described above was applied to the data to find the local $k$-coefficient.

<table>
<thead>
<tr>
<th>ASRB-station</th>
<th>Cloud Cover Observation (CCO)</th>
<th>$k$ (without CCO)</th>
<th>$k$ (with CCO)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Locarno-Monti</td>
<td>y</td>
<td>0.439</td>
<td>0.441</td>
</tr>
<tr>
<td>Payerne</td>
<td>y</td>
<td>0.467</td>
<td>0.461</td>
</tr>
<tr>
<td>Davos</td>
<td>y</td>
<td>0.455</td>
<td>0.444</td>
</tr>
<tr>
<td>Cimetta</td>
<td>n</td>
<td>0.433</td>
<td></td>
</tr>
<tr>
<td>Männlichen</td>
<td>n</td>
<td>0.435</td>
<td></td>
</tr>
<tr>
<td>SLF-Versuchsfeld</td>
<td>n</td>
<td>0.434</td>
<td></td>
</tr>
<tr>
<td>Weissfluhjoch</td>
<td>n</td>
<td>0.430</td>
<td></td>
</tr>
<tr>
<td>Eggishorn</td>
<td>n</td>
<td>0.425</td>
<td></td>
</tr>
<tr>
<td>Les Diablerets</td>
<td>n</td>
<td>0.418</td>
<td></td>
</tr>
<tr>
<td>Gornergrat</td>
<td>n</td>
<td>0.412</td>
<td></td>
</tr>
<tr>
<td>Jungfraujoch</td>
<td>y</td>
<td>0.427</td>
<td>0.426</td>
</tr>
</tbody>
</table>

We used this method to find the $k$-coefficient for the ASRB-stations without cloud observation. For comparison and quality check to the first method we applied this approach also to the stations with cloud observation data. We found a fairly good agreement between the two methods. The thus found
$k$-coefficients have been listed in Table 5.3, and show a slight correlation to the mean water vapor pressure at the station (Figure 5.3).

![Graph](image-url)

**Figure 5.3:** The correlation between the $k$-coefficients and the annual mean ratio of water vapor pressure to temperature shows a linear behaviour with a difference between the stations north and south of the Alps. The offset of DAV and VSF (in comparison to WFJ) may be explained by their location in a valley, resp. in a little depression. $k$-coefficients for other stations could therefore be approximated with this relationship.

### 5.5 The Clear-Sky Index in Practice

It is not easy to find a good data-set to test our clear-sky index algorithm. With the sunshine duration or the direct sun radiation we only know if there is a cloud in front of the sun. Assuming that if there is one cloud, there are others not far away, this data may be compared with the clear-sky index, which is, in contrast, based on the hemispherical longwave downward radiation. The use of the synoptic sky observation to test our algorithm has the disadvantage that it is subjective and that the horizon is weighted too much. Furthermore, this method has the disadvantage of having no or difficult night-time measurements. Since we have no other possibilities we decided to use all these data-sets to demonstrate the usability of the clear-sky index (Figure 5.4 - Figure 5.7).
Detecting Clear-Sky Situations

Figure 5.4: Comparison between the calculated Clear-Sky Index (CSI) and the observed cloud cover for January 1998 at the ASRB-station Davos. Clear-sky situations are defined as CSI>1. All values above indicate cloudy situations and correspond fairly well with cloud amount observation larger than 2/8. The global solar radiation, the longwave downward radiation and the air temperature are plotted for further insight.

Figure 5.5: The observed cloud cover and the calculated Clear-Sky Index for the whole of 1997 at ASRB-station Jungfraujoch. The greater the circles, the more points are included. The observed 3 and 4 eighths cloud cover, which correspond to a CSI>1, may be an overestimation due to clouds at the horizon, which is very wide at this station.
Correlation between the calculated Clear-Sky Index, the measured sky emissivity and the observed cloud cover for 1996 at the ASRB-station Locarno-Monti. The large majority of the observed cloud cover of zero, one or two eighths, corresponds to a CSI≤1 and a low apparent sky emissivity. This good correlation demonstrates the usability of the method.

A comparison between the measured direct solar radiation and the Clear-Sky Index for May 1997 at the ASRB-station Payerne shows a good agreement. Direct solar radiation above 300 W m⁻² always corresponds to a CSI≤1. Calculated night-time values of the Clear-Sky Index are deleted to allow better comparison.

All these examples demonstrate that the clear-sky index algorithm is not perfect but a good instrument to continuously and automatically detect clear-sky situations at different locations and seasons. In conclusion, it can be said that the main advantage of the clear-sky index is its availability in short intervals (depending only on the measurements intervals of the other parameters), the operational usability and its independence from day-light hours.
5.6 Number of Clear-Sky Conditions per Station

The CSI algorithm allows to count and compare the number of clear-sky situations for every ASRB-station. The computation is done based on the 10 minutes data from 1995-1998 and the result is the percentage of 10 minutes clear-sky situations of total (all-sky) situations. Contrary to the sunshine-duration, which does not even depend on clear-sky, the number of clear-sky situations accounts for day and night conditions. Figure 5.8 shows the annual and seasonal means of number of clear-sky situations. The annual means reach the largest values (>35%) at the two stations south of the Alps (Locarno-Monti, Cimetta) and at Gornergrat, which also seems to profit from the often sunny weather in this part of Switzerland. Together with Payerne this three stations also show the largest values (>30%) in summer (July, June, August). All the mountain stations (except Gornergrat) reach only about 20% clear-sky situations, because sunny weather often produces convective clouds in the Alps. In winter (December, January, February) the situation is reversed. All mountain stations (and Locarno-Monti) observe about 40% clear-sky. Payerne, on the contrary, has less than 15% clear-sky situations in winter, because a low stratus cloud layer often dominates this time of the year.

The calculated CSI values shown in Figure 5.8 are the reverse of the cloud amount. High CSI values signify low cloud amount and vice versa. The cloud amount at the ASRB-stations therefore does not show any clear altitude dependence besides the winter and summer differences described above. The differences from station to station are caused by regional anomalies, but are generally small. This result is confirmed by a cloud cover climatology of Switzerland based on cloud cover observation at different altitudes (Kirchhofer, 1995).

![Graph showing annual and seasonal means of number of 10 minutes clear-sky situations in [%] of total (all-sky) situations calculated with the Clear-Sky Index (CSI).

5.7 Summary and Conclusion

The Clear-Sky Index (CSI) demonstrates that the distinction between clear- and cloudy-sky situations can be achieved with the help of longwave downward radiation and the standard meteorological elements air temperature and humidity. The implementation of the CSI offers many new possibilities in the field of meteorology and radiation climatology. Investigations on cloud forcing or the greenhouse effect, such as those in the following chapters, are only possible in this quality and with this amount of data by the use of the CSI.
The CSI allows to monitor the cloud amount continuously (24 h a day), instead to relying on the rare and subjective point measurements of an observer. Furthermore, the weighting of the CSI is much better than synoptic observations because the measured longwave downward radiation is more influenced by clouds at the zenith than at the horizon.

The CSI has no definitive absolute value and could be divided in different subcategories in future applications to define the percentage of cloud coverage, instead of just deciding if it is cloudy (CSI>1) or clear (CSI≤1).

**Figure 5.9:** Partly cloudy-sky at the ASRB-station SLF Versuchsfield. The standard ASRB-instrumentation is on the right, the same two instruments (together with UV-Biometers) for the measurement of the upward fluxes can be seen on the left.
6 Surface Radiation in the Alps

6.1 Introduction

6.1.1 Definition and Relevance
The fact that the different radiation fluxes play an essential role in defining the Earth's climate has long been recognized. The total amount of radiation energy loss or gain can be expressed for every point on earth by the sum of the different fluxes. This sum is called radiation budget (R), which can also be expressed as the sum on the short- (SW\text{net}) and longwave net-radiation (LW\text{net}):

\[ R = (SW \downarrow + SW \uparrow) + (LW \downarrow + LW \uparrow) = SW\text{net} + LW\text{net}. \]  

(6.1)

The earth-leaving (outgoing) fluxes are by definition negative, the incoming fluxes positive. Accurate data of the earth's surface radiation budget (SRB) are urgently needed to improve our understanding of the transfer of internal atmospheric forcing down to the surface, where the radiation budget determines the energy and evaporation budget and thus the energy flux into the atmosphere.

General circulation models (GCMs) and satellite data are essential to understanding the surface radiation budget mechanisms, but they are of limited use until their outputs can be validated against accurately measured radiation data. The radiation budget climatology offers a useful method for understanding the processes of the basic nature of climate and its change.

The SRB depends on the elevation, a fact which influences the dynamics of the atmosphere in regions with mountain chains like the Alps. This dependence stems from the contribution of the water vapor to the overall greenhouse effect of the atmosphere, which strongly decreases with decreasing air temperature and thus with increasing height. The alpine region is therefore an ideal test site for the investigation of the feedback of water vapor in a changing greenhouse scenario.

6.1.2 Measured and Estimated Fluxes
One of the goals of the ASRB project is the investigation of the SRB over a wide range of altitudes. It was not possible to measure all four fluxes at all stations, because the ground below the instruments was not representative for the surrounding environment at many stations.

We measured the whole radiation budget at the three stations Davos, SLF Versuchsfeld and Weissfluhjoch. At Payerne we additionally made use of the upward fluxes from the BSRN-station facility, which were measured with the same type of instruments at the same spot. The missing fluxes at the other stations were estimated based on experiences and investigations from the stations with measured down- and upward fluxes. This was possible because the radiation budget is usually calculated over a certain period (monthly or annual mean), where the spatial short-term variations play a minor role. The meteorological standard elements temperature and humidity were measured at every station and their mean values are explained in the next section.

In the following paragraphs we often talk about annual, summer and winter mean values or gradients. These numbers always stand for the annual mean of the monthly averages, respectively the monthly average of June, July, August (summer mean) and the monthly average of December, January, February (winter mean). The whole calculation is based on 10 minutes data from 1995-1998. For the few
stations with shorter time series we analyzed the data that was available. This concept was chosen instead of the single monthly mean values to reduce the amount of data. With the summer and winter means, the minimum and maximum values are normally located.

We also distinguish between clear- and all-sky mean values. The clear-sky data are extracted with help of the clear-sky index. Clear-sky data are not equally distributed and an observation period of four years (1995-1998) is too short for some stations to find enough clear-sky situations at any time period. The clear-sky mean values were therefore calculated from interpolated data.

6.2 Meteorological Conditions at the Stations

To understand the local differences of the ASRB-stations it is important to know the meteorological conditions at these locations, which are of course influenced by the radiation energy. In Section 5.6 we have already shown the mean cloud amount for every station. The temperature and humidity data give us further knowledge of the different characteristics of the stations.

6.2.1 Air Temperature

Air temperature is measured at all ASRB-stations with a SMI-Thygau about 2 m above the ground. Figure 6.1 shows the annual and seasonal temperature gradient of all ASRB-stations. As theoretically expected the summer lapse rate (0.62°C/100 m) is slightly steeper than the winter lapse rate (0.43°C/100 m) due to frequent inversions and more cloud cover at the lower stations. The means of the three stations Locarno-Monti, Cimetta and Gornergrat are clearly above the values of their neighbouring stations due to the often warmer conditions in the south of Switzerland. In winter the mean temperature of Davos seems to be too low in comparison to the others. This may be due to the valley location, where the temperature in winter often is lower than 100 m above. The opposite is the case in Männlichen and Diablerets, where the winter temperatures seem to be slightly too high. At Diablerets this is probably caused by the sometimes erroneous measurements of the Thygau due to its exposed location (ice accumulation).
6.2.2 Humidity

The humidity at the station is measured as dew point temperature and converted to relative humidity (RH) with the same Thygans at screen level height. For comparisons and usability in the different equations the relative humidity was transformed to water vapor pressure (WV) in [hPa] using the following equations:

\[ WV = \frac{RH}{100} \cdot WV_s, \text{ whereas} \]

\[ WV_s = 6.1121 \cdot \exp \left( \frac{17.502 \cdot t_d}{t_d + 240.97} \right) \]

This empirical equation for the saturated water vapor pressure (WVs) in [hPa] is valid from -30°C to 50°C and is taken from (Buck 1981). The air temperature \( t_d \) has to be in [°C].

![Figure 6.2: Annual and seasonal means of water vapor pressure at all ASRB-stations.](image)

The annual and seasonal means of water vapor pressure are plotted in Figure 6.2. Gradients are similar to the temperatures gradients, with the highest gradient in summer (0.35 hPa) and the lowest in winter (0.14 hPa). A deviation of the southern stations can only be seen at Locarno-Monti and Cimetta in summer, when heavy rainfall is often observed. This high humidity is caused by warm air from the south, which becomes saturated flowing north due to its topographically forced rise. In winter the value of Payerne is quite high due to the often foggy conditions.

6.3 Global Radiation

Global Radiation (SWd) is the sum of the direct solar and diffuse sky radiation. It is the most commonly measured component of the radiation balance. Global radiation depends on latitude, declination, cloud amount, transparency of the atmosphere and surface albedo.

6.3.1 Seasonal Cycle and Altitude Dependence

Daily maximum global radiation (2 minutes averages) can be reached at mountain stations sometimes more than the extraterrestrial solar irradiance (1366 W m\(^{-2}\)), because snow and broken clouds can pro-
duce very high diffuse radiation. It therefore only makes sense to compare the daily maximum for clear sky situations. Figure 6.3 shows all clear-sky 10 minutes values (1995-1998) of the global radiation for the station Gornergrat. The mean daily clear-sky global radiation was calculated by averaging the 10 minutes raw values of global radiation. The often cloudy situations during summer produce some low daily mean values during that time. Nevertheless, we can find a fit to the annual cycle of daily mean global radiation by performing a sine-fit through the daily mean values (Figure 6.3). The annual cycle of the daily maximum global radiation was found with a sine-fit through the daily maxima. The rounded result of this investigation for all ASRB-stations can be seen in Table 6.1. The smallest values are always observed at the beginning of December and the largest in mid June.

![Graph showing annual cycle of daily mean and maximum values of clear-sky global radiation at Weissfluhjoch.](image)

**Figure 6.3**: Annual cycle of daily mean and maximum values of the clear-sky global radiation at the ASRB-station Weissfluhjoch.

**Table 6.1**: Daily maximum clear-sky global radiation (2 minutes averages) for December and June at all ASRB-stations.

<table>
<thead>
<tr>
<th>Daily maximum clear-sky global radiation</th>
<th>December [W m$^{-2}$]</th>
<th>June [W m$^{-2}$]</th>
</tr>
</thead>
<tbody>
<tr>
<td>Locarno-Monti</td>
<td>360</td>
<td>955</td>
</tr>
<tr>
<td>Payerne</td>
<td>360</td>
<td>960</td>
</tr>
<tr>
<td>Davos</td>
<td>385</td>
<td>1055</td>
</tr>
<tr>
<td>Cimetta</td>
<td>385</td>
<td>1055</td>
</tr>
<tr>
<td>Männlichen</td>
<td>390</td>
<td>1060</td>
</tr>
<tr>
<td>SLF Versuchsfeld</td>
<td>400</td>
<td>1080</td>
</tr>
<tr>
<td>Weissfluhjoch</td>
<td>400</td>
<td>1080</td>
</tr>
<tr>
<td>Eggishorn</td>
<td>400</td>
<td>1095</td>
</tr>
<tr>
<td>Diablerets</td>
<td>405</td>
<td>1100</td>
</tr>
<tr>
<td>Gornergrat</td>
<td>410</td>
<td>1110</td>
</tr>
<tr>
<td>Jungfraujoch</td>
<td>420</td>
<td>1130</td>
</tr>
</tbody>
</table>
The annual cycle of the daily mean clear-sky global radiation (from Figure 6.3) at Weissfluhjoch can also be seen in Figure 6.4 in comparison with the annual cycle of daily mean all-sky global radiation. The all-sky global radiation is slightly asymmetric due to higher diffuse radiation in spring (due to higher albedo) than in fall. This causes the maximum of all-sky global radiation already in May, i.e. about one month earlier than the maximum of the clear-sky global radiation. The annual course of the number of 10 minutes clear-sky values per day (# of clear-sky) is also plotted for a better understanding of the difference between the two curves. The correlation between this difference and the number of clear-sky situations is plotted in Figure 6.5.

The dependence of insolation on altitude has mainly been studied under cloud free conditions, where an increase with altitude due to the decrease of the atmosphere’s optical depth has been found (e.g.,
Barry 1981, Muller 1984). For the ASRB-stations this increase of clear-sky global radiation with elevation is shown in Figure 6.6. To establish a relation between insolation and altitude in climatological terms measured monthly mean values of all-sky global radiation have been used. The positive gradients thus found are shown in Figure 6.7. In summer the all-sky gradient is slightly negative below 2000 m a.s.l., positive above 2800 m a.s.l. and zero in between. The negative gradient is caused by enhanced convective activity over the mountain chains leading to an increased cloud shading. The large positive gradient between stations above 2800 m is caused by the glaciers in the vicinity of these stations (Eggishorn, Diablerets, Gornergrat, Jungfraujoch). The glacier surface is always (Jungfraujoch) or often snow covered until mid July. This causes high diffuse solar radiation by multiple reflection between the snow/ice surface and the lower cloud base during convective cloud cover situations. The larger diffuse radiation hence increases the global radiation at these stations. This demonstrates once again the important impact of the snow/ice feedback on the radiation budget.

**Figure 6.6**: Altitude dependence of clear-sky global radiation for annual and seasonal means.

**Figure 6.7**: Altitude dependence of all-sky global radiation for annual and seasonal means. The summer gradient for northern stations below 2000 m a.s.l. is negative, for stations between 2000 and 2800 m zero and for stations above 2800 m positive.
Stations on the south side of the Alps generally measure more shortwave than the ones on the north side. On average, the difference between the northern and southern stations is about 15 W m$^{-2}$, which cannot be explained by the increase of the top of atmosphere irradiance with latitude alone (about 2% per degree). The reason is the difference in the amount of clear-sky situations, which is (for annual means) about 35% on the south side and about 30% on the north side of the Alps (Figure 5.8).

### 6.3.2 Comparison With Other Studies and Models

A similar study with only three stations in Switzerland (Reckenholz 443 m, Rietholzbach 760 m and Arosa 1818 m) was accomplished by Ohmura et al. (1996). They found a summer gradient of -0.8 W m$^{-2}$/100 m, a winter gradient of 1.4 W m$^{-2}$/100 m and an annual gradient of 0.7 W m$^{-2}$/100 m. In order to compare these results with the ASRB-data we investigated the gradient between the stations Payerne (490 m) and Davos (1610 m), which cover about the same elevation (Table 6.2). We found a summer value of -1.6 W m$^{-2}$/100 m, a winter value of 2.5 W m$^{-2}$/100 m and annual value of 1.0 W m$^{-2}$/100 m. The gradient differences between the two studies are significant, although the altitude of the stations are similar. The reason for the deviation may be the different pyranometer generation but is even more likely to be due to the different cloud amount between Payerne and Reckenholz. Nevertheless, both observations and the model found a negative gradient in summer and a positive gradient in winter. The decrease in summer is related to the increased cloud shading caused by enhanced convective activity over the mountain chains. In winter the lowland areas are frequently beneath a stratoform cloud layer formed in stable high pressure situations while the higher altitudes above this layer receive more insolation.

<table>
<thead>
<tr>
<th>[W m$^{-2}$/100m]</th>
<th>annual gradient</th>
<th>winter gradient</th>
<th>summer gradient</th>
<th>annual mean at 1500 m a.s.l.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ohmura et al. (1996)</td>
<td>0.7</td>
<td>1.4</td>
<td>-0.8</td>
<td>131</td>
</tr>
<tr>
<td>GCM nested LAM</td>
<td>-1.8</td>
<td>1.7</td>
<td>-4.2</td>
<td>127</td>
</tr>
<tr>
<td>This study</td>
<td>1.0</td>
<td>2.5</td>
<td>-1.6</td>
<td>153</td>
</tr>
</tbody>
</table>

The same study by Ohmura et al. (1996) also analyzed the annual and seasonal gradient for the north of the Alps by a Local Area Model (LAM), which was nested into a GCM. The annual gradient of the model show a decrease with altitude of -1.8 W m$^{-2}$/100 m. This must be due to the overestimation of the negative summer gradient, which is found to be -4.2 W m$^{-2}$/100 m in the model. This assumption is confirmed by the good agreement between the winter gradient of the model (1.7 W m$^{-2}$/100 m) and the winter gradient of the ASRB-stations (1.4 W m$^{-2}$/100 m). The poor agreements between the annual model gradients and the measurements are not surprising if we compare the absolute values (Table 6.2). The annual mean global radiation at 1500 m a.s.l. is about 20 W m$^{-2}$ too low in Ohmura's measurements and in the model in comparison with the ASRB-mean value.

An older study in the Austrian Alps in Barry (1992) yielded similar gradients for clear-sky global radiation with data from 200 to 3000 m a.s.l as determined with the ASRB-stations. There, the same winter gradient (0.7 W m$^{-2}$/100 m) was found and the summer gradient is only 0.2 W m$^{-2}$ smaller than our gradient (2.6 W m$^{-2}$/100 m) over the whole altitude range.
6.4 Shortwave Reflected Radiation

The shortwave reflected radiation depends on global radiation and the reflectance (albedo) of the surface. At four ASRB-stations shortwave reflected radiation was measured at the same level as the downward fluxes, i.e. about 2-8 m above the ground. The influence of the mast is minimal and has been neglected.

![Graph showing annual cycle of shortwave reflected radiation](image)

**Figure 6.8:** Annual cycle of shortwave reflected radiation at the four stations where the reflected shortwave radiation is measured. The points indicate daily mean values and the curves are the 60 day moving average.

Figure 6.8 shows the measured annual course of the shortwave reflected radiation for the four stations Payerne, Davos, SLF Versuchsfeld and Weissfluhjoch. The curves are based on all-sky daily mean values of the four years 1995-1998. Payerne (490 m a.s.l.) is the only station with an annual curve of shortwave reflected radiation more or less parallel to the global radiation (maximum in summer, minimum in winter). The reason is the low altitude and therefore little snow (i.e. almost constant albedo throughout the year). The other three stations show their annual maximum in spring due to high albedo (snow) and increasing solar declination. The higher stations (SLF Versuchsfeld 2540 m a.s.l. and Weissfluhjoch 2690 m a.s.l.) show their minimum in summer despite high solar elevation, due to very low albedo of the rocky ground. The same minimum level is not observed at the valley station Davos (1610 m a.s.l.) until October due to higher albedo of the surrounding grassland and decreasing sun declination. The significantly higher winter values of Weissfluhjoch in comparison with SLF Versuchsfield can be explained by the large specular reflection of the inclined slopes beyond the station.

The relation between shortwave reflected radiation and global radiation leads to the definition of the albedo \( \alpha = \frac{SW_r}{SW_d} \). The diurnal cycle of albedo is more or less constant for solar elevations above 15°. Because of small quantities of shortwave radiation fluxes and lower pyranometer accuracy for large zenith angles, the diurnal cycle of albedo is only considered for solar elevation angles > 15°.

Figure 6.9 shows diurnal cycles for the same four stations for a perfect clear-sky summer and winter day. The curves are based on the measured 2 minutes-values. Under clear-sky situations the features of the diurnal albedo cycle show some variation. This fact is already described in Plüss (1997) and Konzelmann (1994) and can have various reasons. The range of the variation is generally small. During the clear summer day the stations Payerne and Davos measure a little less during noon than in the morning and afternoon, causing a weak diurnal cycle. The station SLF Versuchsfield shows a slight, but steady increase towards the afternoon. The albedo at the station Weissfluhjoch is very constant.
throughout this summer day. In winter, however, Weissfluhjoch shows a similar diurnal cycle as Davos in summer, with a minimum during noon and maximum after and before sunrise/sunset. The cause is the position of the station on a local peak, where the large specular reflection dependence of snow at low solar elevation is very dominant due to the inclined ground.

Due to the partly pronounced diurnal cycle of albedo, a diurnal albedo value was calculated by averaging an hour before and after solar noon. This method guarantees to consider only the time period with the largest daily solar elevation.

The albedo values for the selected winter day at SLF Versuchsfeld and Davos are relatively low in comparison to the calculated mean winter values (TABLE 6.3). The cause is the large variation of the snow-albedo due to the especially fast aging of snow during sunny conditions.

**FIGURE 6.9: Diurnal cycles of the albedo for a perfect clear-sky day in summer and winter 1998.** The features of the diurnal cycle can change in a small range from station to station and from season to season. The global radiation of Davos (light blue) is shown for comparison.

**FIGURE 6.10: Annual cycle of the albedo for the four ASRB-stations, where the reflected shortwave radiation is measured.** The points indicate daily mean values and the curves are the 30 day moving average.
The mean annual cycle of albedo is shown in Figure 6.10 and was calculated by averaging the daily mean albedo values. Payerne shows only a weak increase during winter time, because clays with snow are rare. The curves for Davos, SLF Versuchsfeld and Weissfluhjoch show a typical annual course with the minimum a few weeks after snow melt, when the ground is saturated with water and a maximum in December/January, when the snow is the driest. The higher winter values of the peak-station Weissfluhjoch in comparison to SLF Versuchsfeld can be explained by the inclined snow covered surfaces below the station, which cause large forward scattering of the direct solar beam.

<table>
<thead>
<tr>
<th>Albedo</th>
<th>Measured</th>
<th>Estimated and used</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Year</td>
<td>Winter</td>
</tr>
<tr>
<td>Locarno-Monti</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Payerne</td>
<td>0.25</td>
<td>0.32</td>
</tr>
<tr>
<td>Davos</td>
<td>0.46</td>
<td>0.79</td>
</tr>
<tr>
<td>Cimetta</td>
<td></td>
<td>0.46</td>
</tr>
<tr>
<td>Männlichen</td>
<td></td>
<td>0.47</td>
</tr>
<tr>
<td>SLF Versuchsfeld</td>
<td>0.66</td>
<td>0.85</td>
</tr>
<tr>
<td>Weissfluhjoch</td>
<td>0.53</td>
<td>0.82</td>
</tr>
<tr>
<td>Eggishorn</td>
<td></td>
<td>0.54</td>
</tr>
<tr>
<td>Diablerets</td>
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<td>0.55</td>
</tr>
<tr>
<td>Gornergrat</td>
<td></td>
<td>0.57</td>
</tr>
<tr>
<td>Jungfraujoch</td>
<td></td>
<td>0.60</td>
</tr>
</tbody>
</table>

The knowledge of the station environment (slopes, glaciers, rocks, grassland) and the altitude (snow coverage) were the two most important components to estimate the albedo for the other stations, where shortwave reflected radiation was not measured. The estimation of the albedo is difficult (uncertainty ca. ±15%) and crucial for the calculation of the net fluxes. The estimated seasonal and annual means of albedo shown in Table 6.3 are based on an area with a radius of about 50 m (estimated view of the instrument on a peak) around the station. Typical albedo values are about 10-20% for rock, 15-30% for grassland and 50-95% for snow.

The differences between the albedo of Weissfluhjoch and SLF Versuchsfeld, which are in close vicinity (500 m distance) of each other, is of special interest. They point out the difference between a typical peak-station (with a partial view down to the valleys) and a typical albedo station in flat terrain. The annual cycles in Figure 6.10 show generally higher albedo values at SLF Versuchsfeld. The slightly smaller winter mean value at Weissfluhjoch can be explained by the presence of a few rocks not covered with snow, which are normally visible from a peak station. The larger annual (0.66) and summer (0.35) mean values of SLF Versuchsfeld are caused by the longer duration of the snow cover, which in general ends about mid May at Weissfluhjoch and at the end of June at SLF-Versuchsfeld. These significant differences show that in mountain regions representative albedo values can not be measured over normally horizontal terrain due to large variation within small areas. The annual and seasonal mean values from SLF Versuchsfeld were therefore adapted to Weissfluhjoch for further altitude dependent investigations.
6.5 Shortwave Net Radiation

Shortwave net radiation ($SW_{\text{net}}$) is the sum of global and shortwave reflected radiation and depends mainly on the solar elevation and surface albedo. The diurnal cycle of $SW_{\text{net}}$ follows the global radiation on a lower level, which is decreased by the shortwave reflected radiation (Figure 6.32). $SW_{\text{net}}$ becomes zero at night and is dominated by the albedo during daytime. In general, this means, the higher the albedo, the lower the shortwave net radiation.

![Figure 6.11: Annual cycle (60 day moving average) of the all-sky (solid lines) and clear-sky (dashed lines) shortwave net radiation for the four stations, where the global and reflected shortwave radiation is measured.](image)

The features of the annual cycle of $SW_{\text{net}}$ are therefore reversed to the one of the albedo. The day to day variation of $SW_{\text{net}}$ is much larger, because the global radiation can vary in a large range from day to day. Figure 6.11 shows the clear- and all-sky annual cycle of $SW_{\text{net}}$ for the four stations where the global and reflected shortwave radiation was measured. The minimum is reached at all stations in December, when the albedo is the highest. The maximum is reached for the two lower stations Payerne and Davos in June. The higher stations reach their maximum in July or August, depending on the duration of the snow cover (albedo). The clear-sky $SW_{\text{net}}$ is always higher than the all-sky $SW_{\text{net}}$ due to higher global radiation under clear-sky conditions.

The difference between clear- and all-sky shortwave net radiation in summer increases with altitude from about 50 to 90 W m$^{-2}$ (Table 6.4). In winter a difference of 20 W m$^{-2}$ can be observed at the two lower stations and only a few W m$^{-2}$ at the mountain stations. Throughout the year the clear-sky $SW_{\text{net}}$ shows about 30 to 40 W m$^{-2}$ higher values than the all-sky $SW_{\text{net}}$.

$SW_{\text{net}}$ can also be described as the amount of shortwave radiation absorbed by the ground. The lowest points in Figure 6.11 indicate $SW_{\text{net}}$ for overcast days. The difference between these points and the clear-sky days show that in summer up to twice the shortwave radiation can be absorbed at clear-sky conditions.

Figure 6.12 shows that there is no obvious altitude dependence of $SW_{\text{net}}$ - the gradient is more or less zero. The two lowland stations (Payerne and Locarno-Monti) register (especially in winter) a little more $SW_{\text{net}}$ due to their generally lower albedo. The $SW_{\text{net}}$ at the mountain stations is more or less constant and very low in winter due to the very high albedo. The summer $SW_{\text{net}}$ is around 180 W m$^{-2}$ and follows the global radiation with higher values at the southern stations due to low cloud amount. The
annual mean values are quite constant around 80 W m$^{-2}$ for the mountain stations and about 110 W m$^{-2}$ for the lowland stations.

**Table 6.4: Annual and seasonal mean values of clear- and all-sky shortwave net radiation for all ASRB-stations.**

<table>
<thead>
<tr>
<th>Station</th>
<th>Year</th>
<th>Winter</th>
<th>Summer</th>
<th>Year</th>
<th>Winter</th>
<th>Summer</th>
</tr>
</thead>
<tbody>
<tr>
<td>LOM</td>
<td>157</td>
<td>71</td>
<td>240</td>
<td>120</td>
<td>47</td>
<td>191</td>
</tr>
<tr>
<td>PAY</td>
<td>159</td>
<td>71</td>
<td>246</td>
<td>108</td>
<td>34</td>
<td>184</td>
</tr>
<tr>
<td>DAV</td>
<td>116</td>
<td>23</td>
<td>250</td>
<td>84</td>
<td>17</td>
<td>170</td>
</tr>
<tr>
<td>CIM</td>
<td>122</td>
<td>25</td>
<td>268</td>
<td>85</td>
<td>18</td>
<td>185</td>
</tr>
<tr>
<td>MAE</td>
<td>121</td>
<td>25</td>
<td>280</td>
<td>82</td>
<td>17</td>
<td>176</td>
</tr>
<tr>
<td>VSF</td>
<td>111</td>
<td>20</td>
<td>287</td>
<td>78</td>
<td>15</td>
<td>176</td>
</tr>
<tr>
<td>WFJ</td>
<td>112</td>
<td>20</td>
<td>288</td>
<td>78</td>
<td>15</td>
<td>176</td>
</tr>
<tr>
<td>EGH</td>
<td>114</td>
<td>21</td>
<td>288</td>
<td>81</td>
<td>16</td>
<td>181</td>
</tr>
<tr>
<td>DIA</td>
<td>113</td>
<td>21</td>
<td>278</td>
<td>78</td>
<td>16</td>
<td>180</td>
</tr>
<tr>
<td>GOR</td>
<td>109</td>
<td>21</td>
<td>283</td>
<td>85</td>
<td>17</td>
<td>208</td>
</tr>
<tr>
<td>JFJ</td>
<td>102</td>
<td>21</td>
<td>268</td>
<td>75</td>
<td>15</td>
<td>188</td>
</tr>
</tbody>
</table>

**Figure 6.12: Altitude profile of all-sky shortwave net radiation for annual and seasonal means. Lower stations show higher values due to lower albedo. Stations with low cloud amount show higher values due to more global radiation.**

### 6.6 Longwave Downward Radiation

Longwave downward radiation ($LW_d$) is emitted and partly absorbed by atmospheric gases. These greenhouse gases are water vapor (the largest contributor), carbon dioxide, ozone, methane, CFC’s, nitrogen compounds and other radiatively active gases. While the effect of reduced atmospheric density with altitude is important for solar radiation, the maximum absorptance by an atmospheric column
under clear sky is only about 25 - 30% of the extra-terrestrial solar radiation. In addition to temperature, longwave radiation fluxes are significantly affected by the increased atmospheric transparency at high elevations due to the decreasing water vapor content. Longwave downward radiation is therefore strongly correlated with humidity and air temperature.

The transmittance spectrum of the atmosphere has a window in the region between 7-14 μm. The earth-atmosphere system loses most of its energy through this window. The presence of clouds almost closes this window. It is therefore worthwhile to specifically investigate the clear-sky characteristics of the longwave downward radiation.

6.6.1 Diurnal Cycle

The diurnal cycle of clear-sky LW↓ is strongly correlated with the diurnal cycle of air temperature and humidity. Figure 6.13 and Figure 6.14 show the summer and winter mean diurnal cycles for all 3 elements at the lowest (Locarno-Monti, 370 m) and highest station (Jungfraujoch, 3580 m) of the ASRB-network. The minimum values of all three elements are observed in the morning just before sunrise. The maximum values appear in the afternoon at about 15:00 UTC. The maximum of LW↓ is normally observed slightly after the temperature maximum, because the heating of the atmosphere is delayed in comparison to the heating of the air temperature at the station.

The range from the daily minimum to the maximum value at Locarno-Monti is about 40 W m⁻² in summer and 20 W m⁻² in winter. At the station Jungfraujoch this range is decreased from 15 W m⁻² in summer to about 7 W m⁻² in winter. This data and investigation with diurnal clear-sky cycles at the other stations showed that the winter range is always about half of the summer range.

In summer two peaks of humidity can be observed at lower stations. The first peak after sunrise is caused by evaporation of the important amounts of ground humidity. Before noon the mixture of the air by convection is so strong that the evaporation at the bottom is not sufficient in order to hold the water vapor pressure. In the afternoon the water vapor pressure increases again (causing the second peak), because the convection is decreasing.
So far we have discussed the clear-sky diurnal cycle; the all-sky daily mean, minimum and maximum cycles of $LW^\downarrow$ are shown for all months at two other stations (Payerne and Eggishorn) in Figure 6.15 and Figure 6.16. The minimum corresponds to the clear-sky diurnal cycle and the maximum to the overcast diurnal cycle. Investigations on all stations demonstrate that the difference between these two extreme conditions within a month is always between 150 and 200 W m$^{-2}$. Smaller differences can generally be observed in the summer months and higher values in the winter months.

In winter the all-sky $LW^\downarrow$ is almost constant throughout the day in Payerne and Eggishorn. In summer the minimum of the all-sky diurnal cycle occurs at about the same time (5:00 UTC) as the minimum of the clear-sky diurnal cycle. The maximum, however, is at about 11:00 UTC (3 hours earlier) at Eggishorn and at about 16:00 UTC (1 hours later than the clear-sky maximum) at Payerne (July and August in Figure 6.15 and Figure 6.16). This displacement of the observed maximum value can be explained at Eggishorn with the development of convective clouds around noon (13:00 local time). At Payerne the one hour displacement can be explained with more clouds in late afternoon due to later onset of convection on the Swiss central plateau.
FIGURE 6.15: Mean, maximum and minimum diurnal cycle of longwave downward radiation for each month at the ASRB-station Payeino. The minimum corresponds to the clear-sky diurnal cycle and the maximum to the overcast diurnal cycle.
Figure 6.16: Mean, maximum and minimum diurnal cycle of longwave downward radiation for each month at the ASRB-station Eggishorn. The minimum corresponds to the clear-sky diurnal cycle and the maximum to the overcast diurnal cycle.
6.6.2 Annual Cycle

Clear-sky annual cycles for $LW_i$, $T$ and $WV$ at the Locarno and Jungfraujoch are shown in Figure 6.17 and Figure 6.18. The 10 minutes raw data were used to calculate a ±10 day average diurnal cycle; i.e. the 10 minutes values of the 10 days before and after one particular day were laid on top of each other to plot one diurnal cycle. This procedure was applied in order to have enough clear-sky points to determine a diurnal cycle. These diurnal cycles were afterwards used to calculate the daily mean values.

The output of this procedure showed better results than the raw data, because clear-sky moments are irregularly distributed with too few points for many of the days. There were still too few points to calculate a reasonable daily mean at some stations with only three years of data (e.g. Jungfraujoch in Figure 6.18). These days were omitted for the calculation of the means, but there were only very few. The daily means of all three elements more or less follow a sine curve. The annual clear-sky minimum
of all three elements appears between the end of November and the end of January. The maximum is observed for all three elements between the end of July and mid-August.

The annual clear-sky cycle in comparison with the all-sky annual cycle at the station Weissfluhjoch is shown in Figure 6.19. The minimum of the all-sky annual cycle surprisingly does not occur until February, i.e., about one month after the clear-sky annual minimum, despite equal or more cloud cover in February. This feature could also be observed at all other stations (Appendix B) and may be caused by the delayed maximum cooling of the atmosphere. The maximum is similar to that for clear-sky between mid-July and mid-August. Table 6.5 shows the annual and seasonal for all stations. The difference between all- and clear-sky values varies depending on station and cloud amount between 10 and 50 W m$^{-2}$. The lower the cloud amount, the smaller the difference (Figure 6.20).
FIGURE 6.19: Annual all- and clear-sky cycle of longwave downward radiation at the ASRB-station Weissfluhjoch.

FIGURE 6.20: The difference between summer means of clear- and all-sky longwave downward radiation shows a fair correlation with the percentage of clear-sky values during the same time. Mountain stations (except Gornergrat) generally have less clear-sky in summer than lower stations due to convective clouds.

The altitude dependence of the all-sky longwave downward radiation is shown in Figure 6.21. The points are almost all on one line except the values for Gornergrat, which are lower due to very low cloud amount. An annual gradient of -2.9 W m\(^{-2}\)/100 m elevation is observed within a range from -2.8 W m\(^{-2}\) in winter to -3.1 W m\(^{-2}\) in summer. This is in accordance with the Stefan-Boltzmann law, which demands for higher gradients with increasing temperatures. The reason for the decrease of LW\(_\downarrow\) is the lower air temperature and, as a result, the smaller water vapor content in the overlying atmospheric column.
### Table 6.5: Annual and seasonal all- and clear-sky mean values of longwave downward radiation at all ASRB-stations.

<table>
<thead>
<tr>
<th>Station</th>
<th>Year</th>
<th>Winter</th>
<th>Summer</th>
<th>Year</th>
<th>Winter</th>
<th>Summer</th>
</tr>
</thead>
<tbody>
<tr>
<td>LOM</td>
<td>280</td>
<td>240</td>
<td>327</td>
<td>311</td>
<td>274</td>
<td>356</td>
</tr>
<tr>
<td>PAY</td>
<td>272</td>
<td>232</td>
<td>320</td>
<td>313</td>
<td>290</td>
<td>347</td>
</tr>
<tr>
<td>DAV</td>
<td>239</td>
<td>204</td>
<td>285</td>
<td>276</td>
<td>243</td>
<td>317</td>
</tr>
<tr>
<td>CIM</td>
<td>238</td>
<td>204</td>
<td>281</td>
<td>274</td>
<td>242</td>
<td>314</td>
</tr>
<tr>
<td>MAE</td>
<td>222</td>
<td>194</td>
<td>258</td>
<td>262</td>
<td>232</td>
<td>300</td>
</tr>
<tr>
<td>VSF</td>
<td>205</td>
<td>175</td>
<td>243</td>
<td>248</td>
<td>218</td>
<td>290</td>
</tr>
<tr>
<td>WEF</td>
<td>204</td>
<td>173</td>
<td>242</td>
<td>248</td>
<td>216</td>
<td>288</td>
</tr>
<tr>
<td>EGH</td>
<td>195</td>
<td>170</td>
<td>229</td>
<td>238</td>
<td>210</td>
<td>275</td>
</tr>
<tr>
<td>DIA</td>
<td>188</td>
<td>165</td>
<td>225</td>
<td>238</td>
<td>205</td>
<td>275</td>
</tr>
<tr>
<td>GOR</td>
<td>186</td>
<td>160</td>
<td>220</td>
<td>225</td>
<td>200</td>
<td>250</td>
</tr>
<tr>
<td>JFJ</td>
<td>173</td>
<td>155</td>
<td>200</td>
<td>226</td>
<td>199</td>
<td>258</td>
</tr>
</tbody>
</table>

**Figure 6.21:** Annual and seasonal mean of the altitude dependence of all-sky longwave downward radiation. Stations on the south side of the Alps generally show lower values due to less cloud amount.

#### 6.6.3 Comparison With Models and Other Studies

There have been very few measurements of longwave downward radiation at mountain stations. Model calculated gradients could therefore hardly be compared with reliable measurements. One of the few investigations comparing simulations and observations is the previously discussed study of Ohmura et al. (1996). They compared three Swiss radiation stations (Reckenholz 443 m, Rietholzbach 760 m and Arosa 1818 m) with the results of a GCM (ECHAM3 T106) nested Local Area Model (LAM). They found an annual mean vertical gradient of -2.7 W m\(^{-2}\)/100 m for the model and -2.8 W m\(^{-2}\)/100 m for the measurements. In order to compare these results with the ASRB-data we investigated the vertical
gradient between the stations Payerne and Davos, which have a similar elevation difference, and observed -3.1 W m\(^{-2}\)/100 m. Having a closer look at the seasons, their model and the observations show a smaller gradient for the summer and a larger for the winter than the annual gradient (TABLE 6.6). The ASRB-data show exactly the same vertical gradient in summer but a larger value in winter (-4.1 W m\(^{-2}\)/100 m). This winter difference may be due to more cloud amount at Payerne than at Reckenholz. It has to be emphasized that both these winter gradients are much too large in comparison with the winter gradient over the whole elevation range (-2.8 W m\(^{-2}\)/100 m).

**TABLE 6.6: Comparison between measured and modelled altitude gradients [W m\(^{-2}\)/100 m] of longwave downward radiation below 1500 m a.s.l. for the north side of the Alps**

<table>
<thead>
<tr>
<th>[W m(^{-2})/100m]</th>
<th>annual gradient</th>
<th>winter gradient</th>
<th>summer gradient</th>
<th>annual mean at 1000 m a.s.l</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ohmura et al. (1996)</td>
<td>-2.8</td>
<td>-3.0</td>
<td>-2.7</td>
<td>309</td>
</tr>
<tr>
<td>GCM nested LAM</td>
<td>-2.7</td>
<td>-3.4</td>
<td>-2.5</td>
<td>296</td>
</tr>
<tr>
<td>This study</td>
<td>-3.1</td>
<td>-4.1</td>
<td>-2.6</td>
<td>296</td>
</tr>
</tbody>
</table>

**FIGURE 6.22: Comparison of annual mean values of longwave downward radiation between measurements from the ASRB-network, three Swiss stations of a different study and a high resolution GCM simulation.**

Ohmura et al. (1996) also observed that the results of the model were always about 15 W m\(^{-2}\) smaller than the observation. They explained this shift with the underestimation of \(\text{LW}\downarrow\) in the radiation scheme of the model. However, the ASRB annual mean values of \(\text{LW}\downarrow\) agree almost perfectly with the model (FIGURE 6.22 and TABLE 6.6). Keeping in mind the deviations of the seasonal gradients between model and observations, the good agreement of the annual means with the model is probably a coincidence. Nevertheless, the pyrgeometer measurements of the ASRB-network are surely better than the delicate and pyranometer measurements used in Ohmura’s study. The highest station (Arosa) of the Ohmura study is only about 14 km distance from Davos, which makes the question even more justifiable. It would be interesting to compare the thus calculated model results with our data over the whole elevation range.
A comparison between our measured clear-sky longwave downward gradients and MODTRAN calculated gradients of the integrated midlatitude summer and winter atmosphere is shown in Figure 6.23. The absolute values of the fluxes and the gradient in the summer atmosphere over the Alps (-3.9 W m\(^{-2}\)/100 m) correspond almost perfectly to the MODTRAN calculated midlatitude summer atmosphere fluxes and gradient (-3.8 W m\(^{-2}\)/100 m). In winter the difference between ASRB fluxes and the MODTRAN calculated midlatitude winter fluxes decrease steadily from about 20 W m\(^{-2}\) at the lower stations to about 3 W m\(^{-2}\) at the highest station. Due to these lower fluxes the MODTRAN gradient (-2.2 W m\(^{-2}\)/100 m) is smaller than the observed ASRB gradient (-2.7 W m\(^{-2}\)/100 m). It can be concluded that the MODTRAN midlatitude summer atmosphere almost perfectly corresponds with the mean summer atmosphere over the Swiss Alps, whereas the MODTRAN midlatitude winter atmosphere is a little bit too cold.

![Figure 6.23: Annual and seasonal altitude dependence of clear-sky longwave downward radiation in comparison with a MODTRAN calculated midlatitude summer and winter atmosphere.](image-url)
Nonetheless, the differences between the SM results and the ASRB measurements are generally small. Especially the summer flux (11 Aug) at the mountain stations is much better simulated with the SM profiles than with the Payerne sonde profile. Exceptions are the lowest and highest stations, which are inexplicably both simulated about 25 W m$^{-2}$ above the measurement. The calculated winter fluxes are close to the measurements for the radiosonde profile and the SM-profiles. The larger differences at the two lowest stations between the model results from the SM-profile and the ASRB measurements indicate that the SM-profile does not correctly simulate the pronounced inversion at these lowland stations.

### 6.7 Longwave Upward Radiation

Longwave upward radiation ($LW^\uparrow$) is the thermal radiation from the surface and depends strongly on the surface temperature ($T_s$) and the emissivity of the ground. It can be obtained by the following equation:

$$LW^\uparrow = \varepsilon \sigma T_s^4 + (1 - \varepsilon) LW^\downarrow,$$  \hspace{1cm} (6.4)
whereby \( \varepsilon_r \) is the emissivity of the surface, \( T_s \) the surface temperature and \( \sigma \) the Stefan-Boltzmann constant. The literature gives emissivities of natural surfaces from 0.9 (dry grassland) to 0.99 (fresh snow). Monthly mean ASRB values showed that \( LW_\text{F} = 0.8 LW^\uparrow \). EQUATION 6.4 can thus be simplified to:

\[
LW_\uparrow = \sigma T_s^4 \cdot \left( \varepsilon_r + (1 - \varepsilon_r) \cdot 0.8 \right). \tag{6.5}
\]

With emissivities between 0.9 and 0.99 longwave upward radiation is hence always between \( 0.980 \sigma T_s^4 \) and \( 0.998 \sigma T_s^4 \). Due to mostly humid surfaces at the ASRB-stations (i.e. \( \varepsilon_s > 0.93 \)), long-wave upward can be calculated with an approximation (maximum error about 2%) using the simple equation:

\[
LW_\uparrow = \sigma T_s^4. \tag{6.6}
\]

The effective surface temperature \( T_e \) is not measured at the ASRB-stations, but it can be calculated from the longwave upward radiation by transformation of EQUATION 6.6. At all stations, where \( LW_\uparrow \) is not measured, it has to be obtained from the 2 m air temperature \( T_a \). The difference between air temperature and surface temperature at the lowland station Payerne and the mountain station Weissfluhjoch is shown in FIGURE 6.25 and FIGURE 6.26. The daily mean difference between the air temperature and surface temperature is negative in summer and positive in winter.

This means that the daily mean surface temperature is warmer than the air temperature in summer and colder in winter. The difference is more pronounced at Weissfluhjoch due to the presence of a snow surface in winter and a rock surface in summer. At Payerne, in contrast, the small differences are caused by the grass surface and the few days with a snow covered ground.

TABLE 6.8: The four stations where longwave upward radiation and the air temperature are measured, show the following annual, summer and winter mean differences between the measured 2 m air temperature and the surface temperature (LW_\uparrow):

<table>
<thead>
<tr>
<th>( T_a - T_s ) [K]</th>
<th>Year</th>
<th>Winter</th>
<th>Summer</th>
</tr>
</thead>
<tbody>
<tr>
<td>Payerne</td>
<td>0.0</td>
<td>0.4</td>
<td>-0.6</td>
</tr>
<tr>
<td>Davos</td>
<td>1.7</td>
<td>6.0</td>
<td>-0.9</td>
</tr>
<tr>
<td>SLF Versuchsfield</td>
<td>2.1</td>
<td>5.7</td>
<td>-1.2</td>
</tr>
<tr>
<td>Weissfluhjoch</td>
<td>0.1</td>
<td>2.8</td>
<td>-3.4</td>
</tr>
</tbody>
</table>

TABLE 6.8 shows the annual and seasonal mean values of the differences between air and surface temperature at the stations where \( LW_\uparrow \) is measured. The annual and winter values of Davos and SLF Versuchsfield are lower than those at Weissfluhjoch, because both stations are located in a small depression, where cold air is often accumulated. On an annual average \( T_a \) is generally not a bad approximation of the \( LW_\uparrow \). Many studies therefore just used \( T_a \) to calculate \( LW_\uparrow \).
Figure 6.25: Monthly mean, maximal and minimal diurnal cycle of the difference between the measured 2 m air temperature and the surface temperature at the ASRB-station Payerne.
FIGURE 6.26: Monthly mean, maximal and minimal diurnal cycle of the difference between the measured 2 m air temperature and the surface temperature at the ASRB-station Weissfluhjoch.
Figure 6.27: Annual and seasonal means of the longwave upward radiation at all ASRB-stations. The values for Payenre, Davos, SLF Versuchsfeld and Weisfluhjoch were measured, all the others were approximated with the help of the air temperature and the knowledge of the local characteristics.

Nevertheless, the estimation of the LW↑ at all other ASRB-stations, where no downward looking pyrgeometer was available, was based on the values shown in Table 6.8. The measured air temperature and the knowledge of the local characteristics was used to estimate the mean T↑. LW↑ could then be calculated with Equation 6.6. The error we made by setting ε↑ to unity can be ignored because it is neutralized by the re-use of Equation 6.6 to calculate LW↑. The approximated LW↑ and its gradient in degree Celsius (for comparison with the air temperature) is plotted in Figure 6.27.

The summer gradient of the surface temperature (-0.53°C/100 m) is smaller than the air temperature gradient (-0.62°C/100 m) in Figure 6.1. This may be due to rocky surfaces at the higher stations, which absorb a lot of global radiation. The winter gradient of the surface temperature (-0.55°C/100 m), in contrast, is slightly larger than the air temperature gradient (-0.43°C/100 m). This may be due to the snow covered ground at the alpine stations, which do not react to air temperature and insolation changes. The annual gradient of the surface temperature (-0.55°C/100 m) is exactly the same as the air temperature gradient.

6.8 Longwave Net Radiation

Longwave net radiation (LW_net) is the sum of longwave incoming and outgoing radiation. Due to lower atmospheric temperature compared to surface temperature, longwave net radiation is generally negative, which corresponds to an energy loss at the surface.

The measured annual cycle of all-sky LW_net shows no pronounced seasonal dependence (cf. Figure 6.28). The all-sky daily mean values of LW_net are highly variable due to the variable cloud cover from day to day. The all-sky annual mean cycle is therefore not very pronounced. The summer mean values are decreased at the mountain stations due to increased cloud amount. The difference between the all- and clear-sky annual cycle therefore strongly depends on the cloud cover. The more clouds, the larger the difference. The clear-sky annual cycle reaches the maximum in summer, the minimum is observed in winter (November to January).
FIGURE 6.28: Annual cycle of longwave net radiation for the four stations where the longwave downward and upward radiation is measured. The solid lines are a 60 day moving average of the daily mean all-sky values. The dashed lines mark the 60 day moving average of the clear-sky longwave net radiation.

FIGURE 6.29: Annual and seasonal mean values of all-sky longwave net radiation at all ASRB-stations.

There is no general agreement in literature about the altitude dependence of the longwave net radiation. The altitude profile through the ASRB-stations shows no clear altitude gradient of LW_{net} for the annual and seasonal mean values (FIGURE 6.29). The annual mean values vary between -50 at Payerne and -80 W m^{-2} at Gornegrat with only minor dependence on altitude. FIGURE 6.30 reveals a weak dependence of LW_{net} on cloud amount. This causes for example the low winter value at Payerne (-30 W m^{-2} due to frequent stratoform clouds) or the high summer value at Gornegrat (-90 W m^{-2} due to few convective clouds at this specific station).
The altitude profile of $LW_{\text{net}}$ without cloud disturbance, i.e., the clear-sky mean values, shows decreasing $LW_{\text{net}}$ with increasing height (Figure 6.31). The steepest gradient is observed in summer with 1.4 W m$^{-2}$/100 m due to a strong decrease of longwave downward radiation with altitude. The lower stations (with a grassland environment) show only a little difference in $LW_{\text{net}}$ between the seasons. The lowest summer value is observed at Payerne (-90 W m$^{-2}$) and the highest at Gornergrat (-135 W m$^{-2}$). The annual mean clear-sky values are about 40 W m$^{-2}$ larger than the corresponding all-sky values and vary between -90 and -120 W m$^{-2}$.

The diurnal cycle of $LW_{\text{net}}$ mainly depends on cloud cover. Very low and thick clouds (fog) can produce longwave net radiation of almost zero. The features of the clear-sky diurnal cycle of $LW_{\text{net}}$ depend predominately on the longwave upward radiation because its variation is generally larger than
the variation of the longwave downward radiation (Figure 6.32). During a clear-sky day the maximum of $LW_{\text{net}}$ is reached after noon, when the ground is heated the most. The daily minimum, in contrast, is observed during night. The difference between the maximum and the minimum clear-sky $LW_{\text{net}}$ value is observed during a clear summer day and can reach values up to $120 \, \text{W m}^{-2}$. These high values are observed at mountain stations due to the strong daytime heating and nighttime cooling of the rocky ground.

### 6.9 Net Radiation

The surface net radiation ($R$) is the sum of shortwave and longwave net radiation and determines the available energy for the non-radiative components (latent and sensible heat) of the surface energy balance. Net radiation is most strongly affected by the absorbed solar radiation $SW \downarrow (1 - \alpha)$ due to the large variation of shortwave net radiation in comparison to the more or less constant values of the longwave emission (longwave net radiation), since

$$R = SW \downarrow (1 - \alpha) + (LW \downarrow + LW \uparrow). \quad (6.7)$$

The diurnal clear-sky cycle has its maximum therefore during solar noon and its minimum during the night. The daily mean values are generally negative in winter and positive in summer (Figure 6.32).

On an annual basis, the longer duration of the snow cover at higher elevations causes the absorbed shortwave radiation to be reduced and, consequently, net radiation tends to decrease with elevation. The annual cycle of net radiation (Figure 6.33) shows exactly these features with lower values at higher stations. In summer, where the maximum is observed, clear-sky net radiation is always more positive than all-sky net radiation due to the higher positive shortwave net radiation. During the minimum in winter however, the negative longwave net radiation overbalances the other components, which causes a higher negative net radiation for clear-sky than for all-sky situations.
The absolute values of net radiation for Alpine locations is poorly known from the observational perspective due to difficult and inaccurate measurements of former investigations. Similarly there is no general agreement about the vertical gradient of net radiation besides the decrease of annual net radiation with altitude. The size of the decrease depends strongly on season due to the changing albedo and cloud amount. The annual mean decrease of net radiation is mainly caused by the lower annual mean solar absorption at higher altitudes due to longer duration of the snow cover.

**Figure 6.34:** Annual and seasonal means of all-sky net radiation at all ASRB-stations.

**Figure 6.33:** Annual cycle of net radiation for the four stations, where all radiation fluxes are measured. The solid lines are a 60 day moving average of the daily mean all-sky values (dots). The dashed lines mark 60 day moving average annual cycle of the clear-sky net radiation.

**Figure 6.34 and Figure 6.35** show that the annual mean of net radiation at lower stations is almost the same for clear- and all-sky situation due to the fact that $SW_{net}$ and $LW_{net}$ both increase by about 40 W m$^{-2}$ for clear-sky situations. All-sky annual mean values are positive for all altitudes and vary between 50 W m$^{-2}$ at the lower stations and 10 W m$^{-2}$ at the highest stations. Clear-sky annual mean values in contrast vary from 50 W m$^{-2}$ at the lower stations to -20 W m$^{-2}$ at the highest stations. The
larger decrease of the clear-sky annual mean values of net radiation (-2.5 W m\(^{-2}\)/100 m) in comparison to the all-sky mean gradient (-1.4 W m\(^{-2}\)/100 m) is caused by the weak altitude dependence of the all-sky LW\(_{\text{net}}\) (-0.3 W m\(^{-2}\)/100 m).

**Figure 6.35:** Annual and seasonal means of clear-sky net radiation at all ASRB-stations.

The summer mean values of clear-sky net radiation are almost constant around 150 W m\(^{-2}\). The highest values have been observed at mountain stations with a rocky vicinity (low albedo), the lowest values at the highest mountains stations, where snow and ice increases the albedo. The summer mean values of all-sky net radiation are decrease slightly from about 120 to 110 W m\(^{-2}\) at stations below 2800 m a.s.l., but are slightly higher (about 120 W m\(^{-2}\)) at the highest stations due to higher absorption caused by multiple reflection between snow and the convective cloud base.

**Figure 6.36:** Summer means of shortwave net, longwave net and net radiation at all ASRB-stations. The net radiation is positive due to high shortwave absorption.

The winter mean values of clear-sky net radiation slightly decrease with altitude from about -30 W m\(^{-2}\) to -90 W m\(^{-2}\). The large decrease between the lowest stations and stations around 1500 m a.s.l. is due to little snow, causing a higher impact of the shortwave net radiation at lowland stations. The winter
mean values of all-sky net radiation show more or less the same features on lower level. At the mountain stations values of about $-50 \text{ W m}^{-2}$ are observed, with slightly lower values at stations with low cloud amount (Gornergrat). At the two lowest stations the inclusion of cloud cover causes a mean winter net radiation of $-16 \text{ W m}^{-2}$ at Locarno and even $+3 \text{ W m}^{-2}$ at Payerne.

In the previous paragraphs we showed that the shortwave net radiation in the Alps generally changes from high values in summer to low values in winter due to high albedo. The longwave net radiation however, is more or less constant throughout the year. This causes a strong overbalance of the shortwave net radiation in summer and a weak overbalance of the longwave net radiation in winter (except at Payerne), which is shown in Figure 6.36 and Figure 6.37.

**Figure 6.37:** Winter means of shortwave net, longwave net and net radiation at all ASRB-stations. The net radiation is negative (except at Payerne) due high albedo causing small shortwave absorption.

**Figure 6.38:** Annual means of shortwave net, longwave net and net radiation at all ASRB-stations. The net radiation is slightly positive due to an overbalance of the shortwave net radiation.
The annual mean values (Figure 6.38) are weakly dominated by shortwave net radiation, which results in small positive values of annual mean net radiation at all stations (decreasing with altitude). The absolute amount of energy over the year available for sensible and latent heat fluxes at the lowest two stations is thus about 4-5 times larger than at the mountain stations.

In summary Figure 6.39 shows the annual mean values of the single radiation fluxes and their altitude dependence. The difference between the four single fluxes gets smaller with increasing altitude due to the increasing shortwave fluxes and the decreasing longwave fluxes with altitude. In contrast to the net fluxes the longwave downward and upward fluxes always overbalance the shortwave global and reflected fluxes. It is very instructive to have a look at the ratio of the single radiation fluxes. The annual mean global radiation is about half of longwave downward radiation at the lower stations, but increases to about two thirds of longwave downward radiation at the highest stations. In combination with Figure 6.38 it can be seen that the ratio of net radiation to longwave downward radiation decreases from about one sixth at the lower stations to about one twelfth at the highest stations. Furthermore, the annual mean absorbed shortwave radiation at the surface (shortwave net radiation) is about one third of the surface emitted longwave radiation (longwave upward radiation) at all altitudes. This demonstrates the efficiency of the greenhouse effect.

6.10 Estimation of Uncertainty

An inventory of possible calibration and measurements errors of the pyrano- and pyrgeometers has already been given in Chapter 2. Possible errors of correction algorithms are described in Chapter 4. Errors induced by parameterization or by estimation of not measured values have not been considered yet. Table 6.9 hence lists the total estimated uncertainty of the different radiation fluxes for the calculated mean values. The increase in accuracy by averaging was calculated by multiplying the total error of a single 2 minutes measurement with \(1/\sqrt{n}\), where \(n\) is the number of measurements.
Table 6.9: Estimated uncertainty for the different radiation fluxes depending on the mean values.

<table>
<thead>
<tr>
<th></th>
<th>10 min. Mean</th>
<th>Seasonal Mean</th>
<th>Annual Mean</th>
</tr>
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<tr>
<td></td>
<td>measured</td>
<td>estimated</td>
<td>measured</td>
</tr>
<tr>
<td>SW↓</td>
<td>1.0%</td>
<td>&lt;0.1%</td>
<td>&lt;0.1%</td>
</tr>
<tr>
<td>SW↑</td>
<td>0.5%</td>
<td>&lt;0.1%</td>
<td>10.0%</td>
</tr>
<tr>
<td>SW_{net}</td>
<td>1.5%</td>
<td>&lt;0.1%</td>
<td>10.0%</td>
</tr>
<tr>
<td>LW↓</td>
<td>1.0%</td>
<td>&lt;0.1%</td>
<td>&lt;0.1%</td>
</tr>
<tr>
<td>LW↑</td>
<td>0.5%</td>
<td>&lt;0.1%</td>
<td>5.0%</td>
</tr>
<tr>
<td>LW_{net}</td>
<td>1.5%</td>
<td>&lt;0.1%</td>
<td>5.0%</td>
</tr>
<tr>
<td>R</td>
<td>3.0%</td>
<td>&lt;0.1%</td>
<td>8.0%</td>
</tr>
</tbody>
</table>

The uncertainty of the estimated annual mean looks quite large. The largest absolute values may cause maximum errors of about 15 W m\(^{-2}\) in the shortwave or longwave radiation fluxes. However, maximum errors of net radiation are much smaller (ca. 5 W m\(^{-2}\)) due to smaller absolute values.

Figure 6.40: Radiation measurement platform on the roof of PMOD/WRC. Permanently shaded instruments on a sun-tracker on the left and the standard ASRB-instrumentation on the right.
Cloud Forcing in the Alps

7.1 Introduction

The ASRB-network allows to study the cloud feedback at different altitudes, which introduces major uncertainties into an understanding of the greenhouse problem. Clouds are a key variable for the balance of the Earth’s radiant energy in regulating the planetary albedo. The increase in planetary albedo cools the Earth-atmosphere system. On the other hand, clouds absorb longwave radiation emitted at relatively high temperatures by the Earth’s surface and emit it back to the Earth. This trapping of longwave radiation by clouds tends to warm the Earth’s surface. Thus clouds have two competing effects on the Earth’s radiation balance: a negative feedback by decreasing shortwave transmission to the Earth’s surface and a positive feedback by increasing longwave re-emission in the atmosphere.

7.1.1 Definition

The effects of clouds are quantified by defining the cloud radiative forcing (CF) (Charlock and Ramanathan 1985). It can be expressed as the sum of the shortwave (CF_{SW}) and longwave (CF_{LW}) cloud radiative forcing:

\[ CF = CF_{SW} + CF_{LW} = (SW_{net} - SW_{net,clr}) + (LW_{net} - LW_{net,clr}). \]

The cloud radiative forcing can be expressed for the top of the atmosphere (T), the Earth’s surface (S) and the atmosphere (A) itself (Gupta et al. 1993). The CF(A) is generally small and can not directly be measured, but can be calculated from CF(T) = CF(A) + CF(S). Surface measurements like the ASRB-data allow to determine cloud radiative forcing at the surface, which in the following is just called cloud radiative forcing (CF) for convenience. The CF is calculated in terms of the all-sky (all) minus the clear-sky (clr) net radiation, which can be distinguished according to CHAPT. 5.

The surface radiation measurements of the ASRB-network allow to determine the altitude dependence of the cloud radiative forcing (or just cloud forcing) at the surface. The important effects of the CF raise questions such as: do clouds, on average, increase or decrease the net radiation? What are the regional and seasonal differences? What is the amount of short- and longwave cloud radiative forcing?

7.1.2 Relevance and Impact

Attempts have been made to answer some of the above questions with the help of General Circulation Models [GCMs] (Cess et al. 1995, Harshvardahn et al. 1990, Cess and Potter 1987). Comparisons between GCM and satellite determined values of cloud forcing showed significant deviations (Gupta et al. 1993, Kiel and Ramanathan 1990). Quantitative GCM intercomparison revealed particularly large differences in the magnitude of warming or cooling due to clouds in different regions. Satellite observations (Wielicki et al. 1995, Ardanuy et al. 1991, Ramanathan et al. 1989) showed that the global mean average CF at the top of the atmosphere is approximately -20 W m\(^{-2}\). This represents a cooling which is about 5 times as large as the predicted warming resulting from direct greenhouse forcing (approximately 4 W m\(^{-2}\)) from a doubled CO\(_2\) concentration. A change in climate can perturb the cloud forcing, which in turn can cause a positive or negative feedback to the initial climate change. Thus, observations of long-term changes in cloud forcing provide insights into the nature of the cloud-climate feedback.
There are only very few surface cloud forcing data available. Moreover, no measurements have been used yet to investigate the altitude dependence of the CF. With regard to a changing climate, it is also important to know the current value of the CF and its impact on the radiation budget.

7.2 Cloud Impact on Radiation Budget

The distinction between the all-sky and clear-sky annual mean values in Chapter 6 has revealed that the clouds decrease or increase the single fluxes of the radiation budget by a significant value (Table 7.1). In contrast to the measured clear-sky diurnal cycle of the different radiation fluxes, Figure 7.1 shows a typical cloudy day in August at the ASRB-station Davos, where the typical regularly featured diurnal cycle of the different fluxes are perturbed by clouds. The air temperature is also represented in the figure to estimate the cloud amount with the help of the difference between the air temperature and longwave downward radiation curve. With this method we can identify one overcast period from 3:00 to 10:30 UTC (with a very short clearing around 6:00 UTC) and another from 20:00 to midnight (with short clearing around 23:00 UTC). Global radiation indicates that after 10:30 UTC the coherent cloud cover changed to broken drifting clouds, which shade and un-shade the direct beam of the sun in short time periods.

The clouds cause global, shortwave reflected and shortwave net radiation to become very small. The impact on the longwave fluxes is an increase of the longwave downward and a decrease of the longwave net radiation against zero. These two effects affect the net radiation significantly. The presence of low and thick clouds causes the net radiation to become almost zero. During nighttime the net radiation follows the longwave net radiation due to the absence of shortwave fluxes and is therefore slightly negative under a cloudy sky. During daytime net radiation follows the features of the global radiation and is therefore slightly positive due to the overbalance of the shortwave net radiation. The change in net radiation caused by clouds for this summer day in Davos is thus about 100 W m$^{-2}$ during night and about 600 W m$^{-2}$ during day (Figure 7.1).
### 7.3 Local and Seasonal Differences

#### 7.3.1 Shortwave Cloud Forcing

Clouds generally have a much larger albedo than the earth surface, especially over oceans. This causes large absorption of solar radiation (large shortwave net radiation) at the ground during clear-sky conditions. Clouds cool the earth-atmosphere system by reflecting incoming solar radiation back to space. Shortwave cloud forcing \((CF_{SW})\) is therefore always negative.

The annual cycle of shortwave cloud forcing at one lowland (Payerne) and one mountain station (Weissfluhjoch) shows a large seasonal difference at the mountain station due to convective cloud cover (Figure 7.2). The impact of the frequently present stratoform cloud layer at Payerne during winter can probably not be observed due to low solar declination. The day to day variation is large at both stations due to fast changing conditions in Switzerland and only limited data series (2-4 years), from which daily means were calculated.

The cloud forcing in summer shows that the amount of the cooling is large (about \(-110 \text{ W m}^{-2}\)) at mountain stations with a rocky vicinity due to the large difference between the ground- and cloud surface albedo. However, the highest stations show lower values of \(CF_{SW}\) due to the similar albedo of the partly snow covered vicinity and the cloud surface albedo (Figure 7.3). In winter shortwave cloud

| Difference between all-sky minus clear-sky (average of annual mean of all ASRB-stations) |
|----------------------|----------------------|
| SW \(\downarrow\)     | -60 W m\(^{-2}\)     |
| LW \(\downarrow\)     | 40 W m\(^{-2}\)     |
| SWnet                | -35 W m\(^{-2}\)     |
| LWnet                | 40 W m\(^{-2}\)     |

**Table 7.1:** Impact of cloud cover on the annual mean values (from Chap. 6) of selected radiation fluxes.
forcing is very low and constant (about \(-5 \text{ W m}^{-2}\)) at mountain stations due to the snow cover, which causes solar absorption to be very small. At the lowest two stations this seasonal difference is small due to the smaller variation of the annual albedo cycle. The annual mean \(CF_{SW}\) decreases weakly with altitude from about \(-50 \text{ W m}^{-2}\) at the lower stations to \(-20 \text{ W m}^{-2}\) at the higher stations, causing a gradient of \(0.51 \text{ W m}^{-2}/100 \text{ m}\).

**FIGURE 7.3**: Annual and seasonal mean shortwave cloud forcing and its altitude dependence.

### 7.3.2 Longwave Cloud Forcing

Longwave cloud forcing \((CF_{LW})\) is always positive because the trapping of the longwave radiation by clouds warms the earth-atmosphere system. The annual cycle of longwave cloud forcing at one lowland (Payerne) and one mountain station (Weissfluhjoch) shows no large seasonal differences in comparison to the shortwave cloud forcing (FIGURE 7.4). Nevertheless, Weissfluhjoch yields higher values...
in summer than in winter and vice versa in Payerne. The effect of the stratoform cloud layer in Payerne and clear-sky at Weissfluhjoch is clearly visible from mid January to mid February. Longwave cloud forcing seems to be more sensitive to cloud amount than the shortwave cloud forcing because it is not dependent on insolation. There may still be some dependence to insolation due to the sensitivity of longwave upward radiation on insolation (Figure 6.27). The day to day variation of $CF_{LW}$ is large due to the reasons already discussed in the previous section.

Investigations on all stations revealed that the summer mean values of $CF_{LW}$ shows an increasing trend with altitude from 30 W m$^{-2}$ at the lowest stations to 65 W m$^{-2}$ at the highest station (Figure 7.5) due to the increasing cloud amount with altitude. The winter mean values are almost constant around 45 W m$^{-2}$ for all altitudes (except Payerne, Figure 7.5). The generally higher summer values are caused by the larger longwave absorption (longwave net radiation) in summer. Annual mean values hence show a slight increase with altitude of about 0.58 W m$^{-2}$/100 m from about 35 W m$^{-2}$ at the lowest station to 60 W m$^{-2}$ at the highest station.

The impact of the clouds can be observed particularly well at Payerne and Gornergrat. The station Payerne registers its highest values in winter due to large amounts of cloud and despite low longwave net radiation. The station Gornergrat, in contrast, shows a very low summer value due to small amounts of cloud and despite high longwave net radiation. However, not every feature can be explained with cloud cover. For example, the winter value of $CF_{LW}$ at Locarno-Monti is surprisingly higher than the summer value despite lower cloud amount in winter (Figure 5.8). Other elements, which can influence cloud forcing are therefore discussed in Section 7.4.

The difference between the summer and winter mean values of the longwave cloud forcing at mountain stations are relatively small in comparison to the shortwave cloud forcing differences due to small change between the summer and winter longwave net radiation.

### 7.3.3 Net Cloud Forcing

Net cloud forcing ($CF_{Net}$) is the sum of the shortwave and longwave cloud forcing and determines the total impact of the clouds on the earth-atmosphere system. The seasonal differences are shown in Figure 7.6 for one lowland (Payerne) and one mountain station (Weissfluhjoch). The annual cycle of
net cloud forcing is not very pronounced, but shows positive values in winter and negative values in summer for both stations, indicating a cooling effect by clouds in summer and a warming effect in winter.

**Figure 7.6:** Mean daily values (dots) and monthly smoothed mean annual cycle (solid line) of net cloud forcing at one lowland and one mountain station of the ASRB-network.

**Figure 7.7:** Annual and seasonal mean net cloud forcing and its altitude dependence at all ASRB-stations.

Figure 7.7 shows the annual and seasonal means of the net cloud forcing for all ASRB-stations. The winter mean values show a weak increase with altitude and are positive due to the very low shortwave cloud forcing. The southern station Locarno-Monti, Cimetta and Gornergrat show slightly lower values, although the winter mean cloud amount is not particularly low at these stations. This could be an indication that temperature and/or humidity, which are both a little higher at these three stations, have some impact on the cloud forcing (SECTION 7.4). The summer mean values of the $CF_{Net}$ are negative and show more or less the same altitude dependence as the $CF_{SW}$ due to overbalance of the $CF_{SW}$. The
absolute values of the positive summer net cloud forcing are similar to the negative winter net cloud forcing. The annual mean values show an altitude dependence (1.1 W m\(^{-2}\)/100 m) of net cloud forcing, which ranges from -10 W m\(^{-2}\) at Payerne to 31 W m\(^{-2}\) at Jungfraujoch. This result reveals that the total impact of clouds is slightly negative at the lowland stations and positive in the alpine regions. This means that clouds reduce the effect of altitude by cooling the lower stations and warming the mountain ones. This is also demonstrated by the all-sky altitude gradient (-1.4 W m\(^{-2}\)/100 m) and clear-sky altitude gradient (-2.5 W m\(^{-2}\)/100 m) of net radiation in Section 6.9.

7.4 Discussion

Our investigations on the one hand and sensitivity studies by Gupta et al. (1993) on the other hand showed that \(CF_{SW}\) mainly depends on the available insolation, surface albedo and cloud transmittance. Furthermore \(CF_{LW}\) mainly depends on cloud bottom height and water vapor content of the atmosphere. It is therefore not astonishing that the correlation coefficient between the annual cloud amount and the \(CF_{SW}\) and \(CF_{LW}\) is low (\(R^2 = 0.15\), resp. 0.11) for the ASRB-network. In comparison to the small cloud cover variations, the impact of the other parameters such as insolation, albedo and water vapor are too large within Switzerland, due to complex topography and the large altitude range. The sensitivity of the cloud forcing to the above parameters also helps to understand the features of its annual and seasonal mean values. In Figure 7.3 and Figure 7.5, for example, we observe an altitude dependence of the annual \(CF_{SW}\) and \(CF_{LW}\), although annual mean cloud amount shows no altitude dependence (Section 5.6).

![Figure 7.8: Scatterplots between annual mean shortwave cloud forcing, albedo and global radiation for all ASRB-stations.](image)

The slight decrease of the annual shortwave cloud forcing with altitude indicates that the impact of clouds is weaker at higher stations. This fact can be explained by the generally higher insolation and surface albedo during more than 6 months at high mountain stations. The correlations between these parameters are shown in Figure 7.8. Locarno-Monti is far away from the linear fit due to its little cloud cover caused by its location on the south side of the Alps.
The increase of annual $CF_{LW}$ with altitude indicates on the contrary that the impact of clouds is stronger at higher stations. This can be explained by the fact that $LW$ is strongly dependent on water vapor content and temperature of the first 1000 meters above the instrument (Chapter 8). The smaller distance between cloud base and station at high altitudes is therefore mainly responsible for the large impact of clouds at higher stations. Although water vapor was measured as station humidity and not as the water column above the station, it yields a good correlation (Figure 7.9).

The good correlation of shortwave cloud forcing with insolation and albedo and of $CF_{LW}$ with water vapor and air temperature not only reveals the direct cause for the altitude dependence of the short- and longwave cloud forcing, but also the altitude dependence of the net cloud forcing.

Nevertheless, the impact of the cloud cover can be observed in summer, when the albedo and insolation differences are small and not altitude dependent. The results of the correlation of short- and longwave cloud forcing with cloud amount (Clear-Sky Index) are shown in Figure 7.10. The lower stations show a high Clear-Sky Index (CSI) and a corresponding small $CF_{SW}$ and $CF_{LW}$. The opposite can be observed at the higher stations (with the exception of the two highest stations Jungfraujoch and Gommergrat) with low CSI (frequent clouds) and large $CF_{SW}$ and $CF_{LW}$.

Short- and longwave cloud forcing are differently affected by clouds. In contrast to the domination of the longwave cloud forcing in winter and the domination of the shortwave cloud forcing in summer, the annual mean values almost cancel each other, especially at the low altitude stations (Figure 7.11). However, longwave cloud forcing dominates above 2000 m a.s.l., causing a positive net cloud forcing at mountain stations.

This interplay of $CF_{SW}$ and $CF_{LW}$ and the resulting net cloud forcing is shown in summary in Figure 7.12 for the annual and seasonal means of all ASRB-stations. The winter mean values are fairly constant with the exception of Payere and Locarno-Monti, where short- and longwave cloud forcing is increased due to frequent stratoform cloud cover. The summer mean values show a clear dependence on the distribution of cloud amount, especially for the shortwave cloud forcing. The seasonal difference of net cloud forcing is small for all stations, except for Jungfraujoch due to relatively large $CF_{SW}$ in summer. The annual mean values yield the already discussed altitude gradient of net cloud forcing due to its dependence on water vapor, albedo and insolation.
With the aim of connecting our cloud forcing values with other data, we compared the ASRB-data with satellite derived northern hemisphere surface cloud forcing values from Gupta et al. (1993). In the assumption that the altitude range may simulate the latitude range, we averaged the annual cloud forcing values of the three lowest ASRB-stations (TABLE 7.2). The smaller shortwave value and the larger longwave value in Gupta’s study show either the problems (inaccuracy) in the determination of satellite derived surface radiation fluxes, the impact of the large sea surface, or probably even more the difficulty of such a comparison. The ASRB-data show a cancellation of the short- and longwave cloud forcing, which is a midlatitude speciality. Gupta’s satellite data demonstrate positive $CF_{Net}$ higher latitudes and negative $CF_{Net}$ at lower latitudes, which is in agreement with the positive values of $CF_{Net}$ at the higher ASRB-stations and the negative values at the lowest two stations.
TABLE 7.2: Comparison of ASRB cloud forcing values and satellite derived northern hemisphere data.

<table>
<thead>
<tr>
<th></th>
<th>Satellite derived N.H. data (Gupta et al., 1993)</th>
<th>Average of the three lowest ASRB-stations</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Annual means [W m⁻²]</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>$CF_{SW}$</td>
<td>-47.6</td>
<td>-40.3</td>
</tr>
<tr>
<td>$CF_{LW}$</td>
<td>31.8</td>
<td>40.0</td>
</tr>
<tr>
<td>$CF_{Net}$</td>
<td>-15.8</td>
<td>-0.3</td>
</tr>
</tbody>
</table>

A common misconception is that because clouds globally cool the present climate, they will act likewise to moderate global warming. It is, however, the change in $CF_{Net}$ associated with a change in climate, that governs cloud feedback. An increasing greenhouse-effect, for example, induces a change in
climate by direct radiative forcing. Thus, if \( \Delta CF_{\text{Xel}} \) has the same magnitude as the direct forcing, so that the cloud feedback is positive, the warming is amplified by a factor of 2.

The results shown demonstrate that we can successfully determine the shortwave, longwave and net cloud forcing at the surface from radiation budget measurements and the help of the developed clear-sky index, which allows to distinguish between clear-sky and all-sky conditions. Observations of long-term changes of cloud forcing are important for cloud-climate feedback monitoring and can be used to provide confirmation of previously untestable results of climate models.

**Figure 7.13:** Travelling Standard at the ASRB-station Diablerets.
8 Longwave Downward Radiation and Greenhouse Effect

8.1 Introduction

It is desirable to identify the future development of the global climate by observations. One such possibility is the use of an early detection of a radiation change as a means of climate prediction. The future development of the longwave atmospheric radiation at the earth’s surface is of particular interest as this component is closely related to warming.

Global climate models show an immediate response in longwave downward radiation of only about 1.2 W m\(^{-2}\) from a doubled CO\(_2\)-concentration due to direct forcing mechanisms. However, they also show that there is an additional increase of about 10 to 30 W m\(^{-2}\) due to increased water vapor content due to increased surface temperature (Ramanathan 1985).

The importance and magnitude of longwave downward radiation has generally not been perceived because, climate models, and in particular GCMs, only analyze net flux at the surface. The change in the net flux is signiﬁcantly smaller than the change in the individual components.

Recent transient experiments (Hansen et al. 1988, Manabe et al. 1991, and Cubasch et al. 1992) indicate that a detectable increase in the longwave downward radiation will precede those of other climatological elements. The air temperature increase will be signiﬁcantly delayed due to heating of the oceans and the precipitation change will be diﬃcult to detect because of its large year to year variation (personal communication A. Ohmura). The longwave atmospheric radiation has also been shown to have one of the smallest year to year variations (1% standard deviation with respect to the annual mean at the ASRB-stations) and is not as location dependent as the air temperature measurement. With the development of improved calibration and measurement techniques in the early phase of the ASRB project the measurement accuracy of the longwave radiation was improved by a factor of 5 from approximately ±15 W m\(^{-2}\) to ±3 W m\(^{-2}\) (CHAPT. 2). This improvement makes the longwave downward radiation one of the most promising elements for an early direct detection of the increasing greenhouse effect. One of the long-term goals of the ASRB-project is hence the early detection of the atmospheric radiation increase by a forced greenhouse effect.

8.2 Definition of the Greenhouse Effect

The greenhouse effect (or greenhouse radiative flux) \(G\) is popularly referred to as the energy trapped in the atmosphere and is defined as the difference between the thermal radiation emitted by the surface \(LW_{\text{Surface}}\) and that escaping to space at the top of the atmosphere \(LW_{\text{TOA}}\)

\[
G = LW_{\text{Surface}} - LW_{\text{TOA}}.
\]  

Equation 8.1 demonstrates that the greenhouse radiative flux \(G\) is unfortunately not directly measurable from the Earth’s surface. This chapter will show that the greenhouse effect and the longwave downward radiation are strongly correlated and how the former can be determined experimentally by measuring the latter.
8.3 Origin of Longwave Downward Radiation

In Chapter 6, Figure 6.13 we already showed the close connection between the diurnal cycle of longwave downward radiation, station temperature and humidity. This strong dependence between these quantities allows the parametrization of clear-sky longwave downward radiation using only the station temperature and humidity. The success of this method has already been described in publications by Angström (1918) or Brunt (1932). The main reason is the strong dependence of the atmospheric emittance on the total water vapor content, which mainly determines the amount lost through the atmospheric window. This implies a wider atmospheric window for higher stations, i.e. a larger difference between the longwave radiation and the corresponding Planck curve. At low altitudes, in contrast, the atmospheric window is relatively small, preventing a large impact of gases emitting at higher altitudes.

The humidity at the station is related to the total water vapor content above the station. The water vapor content is connected to air temperature through the Clausius-Clapeyron law (exponential decrease of saturation vapor pressure with decreasing temperature). Therefore, the longwave downward radiation decreases with increasing altitude due to rapidly decreasing water vapor content with height.

Figure 8.1 demonstrates the distribution of water vapor in the atmosphere and the corresponding longwave downward flux based on a radiosonde profile from Payerne on August 11 1998, which was a perfectly clear day. MODTRAN calculations were performed for two different heights (corresponding to Payerne and Jungfraujoch) to determine the origin of the longwave downward radiation. The result for the humidity reveals that almost 50% of the water vapor is within the first 1000 m above a station. This is the main cause that about 90% of the longwave downward radiation originates from the first 1000 m above the station. Two thirds of longwave downward radiation even originates from the first 100 m above a station. Calculations for Payerne (490 m a.s.l.) and Jungfraujoch (3580 m a.s.l.) only showed small differences in the origin of the longwave downward flux within the altitude range of the ASRB stations. The largest differences were observed within layers closest to the station. It is very instructive to see that 40% of longwave downward radiation at 490 m a.s.l. and 30% at 3580 m a.s.l. still originates from the first 10 m above the station. A calculation for a virtual station at 10 000 m a.s.l. was also performed and yields lower percentages than expected, but 75% of longwave downward radi-
Longwave Downward Radiation and Greenhouse Effect

...portion still originates from the first 1000 m above the virtual station. These results demonstrate that longwave downward radiation is strongly dependent on humidity and therefore on air temperature measured at the station.

8.4 Impact of Atmospheric Water Vapor

![Figure 8.2: Altitude dependence of annual mean clear-sky temperature normalized longwave downward radiation, which is caused by the decreasing water vapor content with increasing elevation.](image)

Water vapor in the atmosphere is responsible for more than half of the earth's greenhouse effect, and therefore is a major contributor to the problem of global warming if the amount of water vapor in the atmosphere increases. Hence, it is important to investigate the greenhouse radiative flux and the corresponding longwave downward flux as a function of water vapor.

In the previous section we showed that longwave downward radiation is mainly dependent on air temperature and humidity at the station. To exclude the effect of local temperature, longwave downward radiation can be normalized to the Planck radiation with the station temperature. Figure 8.2 shows the normalized annual mean clear-sky longwave downward radiation at all ASRB-stations. As expected the normalized longwave radiation decreases with increasing altitude. The generally lower values of the southern stations (Locarno-Monti, Cimetta, Gornergrat) may indicate that during clear-sky conditions the atmosphere on the southern slope of the Alps is somewhat drier.

The strong dependence of water vapor on temperature leads to a strong feedback mechanism. Consider an initial increase in the global temperature because of direct infrared trapping resulting from, say, an increase in carbon dioxide. As the surface and atmosphere warm, the saturation vapor pressure increases exponentially with temperature, and more water stays in the atmosphere. This additional moisture then adds to the total absorptance of the atmosphere by trapping more infrared radiation and driving the temperature even higher. This feedback can amplify the original perturbation by as much as a factor of two (Hansen et al. 1984). The correlations between temperature, water vapor, greenhouse flux and longwave downward radiation are discussed in Section 8.5 and Section 8.6.

Recently, it has been noted that there may already have been an increase of over 5% in the atmospheric water vapor content in the northern hemisphere during this century (Elliot 1998). It is therefore essen-
tial to accurately quantify the increase in the greenhouse flux due to the greater abundance of water vapor. We have used MODTRAN to calculate the clear-sky radiative greenhouse flux and the corresponding longwave downward flux associated with such an increase of atmospheric water vapor. The water vapor was thus changed by 5% at all heights, which implicates a higher absolute change at low altitude. This kind of water content change was chosen because it corresponds to observed seasonal changes (Shine and Sinha 1991).

**Table 8.1: Impact on longwave downward radiation (LW↓) and on radiative greenhouse flux (G) due to changes of water vapor content and CO₂ during the last 100 years. The computations were performed with MODTRAN for different altitudes in a clear-sky midlatitude summer (MLS) and a subarctic winter atmosphere (SAW).**

<table>
<thead>
<tr>
<th>[W m⁻²]</th>
<th>5% increase of atmospheric water vapor</th>
<th>Increase of CO₂ from 295 to 360 ppm</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>ΔLW↓ (MLS)</td>
<td>ΔG (MLS)</td>
</tr>
<tr>
<td>500 m</td>
<td>2.62</td>
<td>0.68</td>
</tr>
<tr>
<td>1500 m</td>
<td>1.83</td>
<td>0.75</td>
</tr>
<tr>
<td>2500 m</td>
<td>1.33</td>
<td>0.74</td>
</tr>
<tr>
<td>3500 m</td>
<td>1.09</td>
<td>0.69</td>
</tr>
</tbody>
</table>

Table 8.1 shows the results of this water content change (all other elements were held constant) for a midlatitude summer (MLS) and a subarctic winter atmosphere (SAW) at four different altitudes. To quantify and compare this forcing, the change in the greenhouse flux and the corresponding longwave downward flux was also calculated for the direct forcing of the observed CO₂ increase (without other greenhouse gases and all other elements were held constant) during the last 100 years (295-360 ppm). The calculated impact of the probable water vapor increase on longwave downward and greenhouse flux yields larger values than the impact of the observed CO₂ increase. A further amplification caused by water vapor is the enhanced absorption of solar radiation in the atmosphere, which also leads to an additional heating of the troposphere. These two processes of the water vapor feedback may be responsible for an increase in longwave downward radiation of almost 4 W m⁻² during the last century. The increase of LW↓ is smaller at higher altitudes as theoretically expected and confirmed by measurements (Figure 8.2) due to the water vapor feedback.

The impact of the CO₂ increase on LW↓ in lower altitudes during the same period is comparatively small, but shows a clear increase with altitude, due to less water vapor. This trend is caused by a larger atmospheric window resulting from less overlapping absorption bands between CO₂ and H₂O at higher altitudes (Houghton et al. 1995). The impact of CO₂ is thus larger in a drier atmosphere. It can therefore be concluded that a possible increase of longwave downward radiation is best measured at dry locations like high altitude stations. This idea is strengthened by the result of some climate models, which forecast the largest temperature increase at midlatitude at about 500 hPa (Manabe et al 1991).

Both forcings (the water content and CO₂ change) have a relatively low impact on the greenhouse radiative flux, but an enhanced impact on longwave downward radiation, especially the water vapor feedback. Due to such feedback mechanisms and the above mentioned climate model results, it is in any case important to measure longwave downward radiation at different altitudes.
8.5 Correlations With Temperature and Humidity

Due to the fact that the greenhouse effect can not directly be measured from the earth surface, MODTRAN has been used to analyze the altitude dependence for clear-sky conditions - but not without verifying the underlying processes with measurements. In a first step the ASRB-data were used to calculate correlations between longwave downward radiation, air temperature and precipitable water. In a second step correlations were obtained with the same parameters from the model output results.

8.5.1 Correlation With Measured Data

Figure 8.3 shows the correlation of the daily mean clear-sky values of surface temperature (=air temperature) and the precipitable water at four different ASRB-stations, each about 1000 m higher than the other. The precipitable water ($PW$) of the column was calculated from the relative humidity ($RH$) at the station with the empirical formula of Leckner (1978),

$$ PW = 0.493 \cdot RH \cdot \frac{(WV_s)}{T_a}, \quad (8.2) $$

whereas the water vapor pressure in saturated air ($WV_s$) is given by Equation 6.3. The logarithm of the precipitable water is plotted to account for the exponential dependence of water vapor pressure on temperature. The linear fits through the data show an altitude dependence of increasing PW with increasing temperature as expected. The slope of the Jungfraujoch-data seems to be a little bit too flat for the highest station in comparison to the one of the Weissfluhjoch data. The steeper slope of the station Gornergrat ($d\ln PW/dT=9.42*10^{-2}$, not shown), which is between Weissfluhjoch and Jungfraujoch confirms this observation. The cause might be the extreme location of Jungfraujoch on an exposed saddle between a glacier basin and a huge north facing rock wall, which possibly influences cloud development by local condensation. These extreme conditions also affect the Thygan with snow and ice, which causes erroneous measurement of temperature and humidity.

![Figure 8.3: Daily mean clear-sky values of precipitable water versus temperature at four ASRB-stations.](image)

Figure 8.4 shows the significant increase of normalized longwave downward radiation ($nLW$) with air temperature on the one hand and with precipitable water on the other. The steeper slopes at the two higher stations (on the left graph) reveal that longwave downward radiation seems to be more sensitive to temperature changes at higher altitudes. Again the slope of the Jungfraujoch-data is too low in com-
comparison to Weissfluhjoch and Gornegrat ($dnLW/dT=6.43*10^{-2}$, not shown). The precipitable water (on the right graph) does not yield this behavior due to the water vapor feedback, which is larger at lower stations.

**FIGURE 8.4:** Daily mean clear-sky values of precipitable water and temperature versus normalized longwave downward radiation at four ASRB-stations.

### 8.5.2 Correlation With Modelled Parameters

![Figure 8.5](image)

**FIGURE 8.5:** MODTRAN calculations of clear-sky temperature versus precipitable water for four different atmospheres and altitudes.

The same correlations are now calculated with MODTRAN outputs of temperature, precipitable water and longwave downward radiation for four different atmospheres at four different altitudes corresponding more or less to the station elevations of the previous section. The selected MODTRAN integrated atmospheres are the midlatitude summer (MLS), midlatitude winter (MLW), subarctic summer (SAS) and subarctic winter (SAW) atmospheres. The CO$_2$-concentration was set to 360 ppm and the ozone-concentration was kept at the MLW-value.
Figure 8.5 shows the modelled temperature versus precipitable water and corresponds to Figure 8.3 in the previous section. The intercepts and slopes yield about the same absolute values as the measured values. The higher sensitivity of water vapor at higher altitudes with increasing temperature is a little less pronounced but generally agrees well.

![Graph showing temperature versus precipitable water](image)

**Figure 8.6:** MODTRAN calculation of clear-sky precipitable water and temperature versus normalized longwave downward radiation for four different atmospheres and altitudes.

**Table 8.2:** MODTRAN output and calculated greenhouse effect for the midlatitude summer atmosphere at four altitudes.

<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>500 m</td>
<td>325</td>
<td>292</td>
<td>17.56</td>
<td>269</td>
<td>412</td>
<td>143</td>
</tr>
<tr>
<td>1500 m</td>
<td>284</td>
<td>287</td>
<td>9.39</td>
<td>262</td>
<td>387</td>
<td>125</td>
</tr>
<tr>
<td>2500 m</td>
<td>246</td>
<td>282</td>
<td>4.77</td>
<td>253</td>
<td>360</td>
<td>107</td>
</tr>
<tr>
<td>3500 m</td>
<td>210</td>
<td>276</td>
<td>2.40</td>
<td>242</td>
<td>330</td>
<td>88</td>
</tr>
</tbody>
</table>

Figure 8.6 shows the modelled temperature and precipitable water versus normalized longwave downward radiation (nLW↓). The intercepts and slopes from both graphs are again very similar to the one in Figure 8.4 in the previous section. The rate of nLW↓ increase with temperature and water vapor is slightly too weak in the linear fits of the model values in comparison to the measured data. Nevertheless, they confirm the result of the measured ASRB-data, i.e. that longwave downward radiation increases linearly with increasing temperature and precipitable water. The modelled values also show a larger increase of longwave downward radiation with increasing temperature at higher altitudes. Corresponding to the measured values, the modelled points of longwave downward radiation show no clear altitude dependence with increasing precipitable water. The model output parameters and the calculated greenhouse effect for the midlatitude summer atmosphere are shown in Table 8.2.
8.6 Correlations With the Greenhouse Effect

The greenhouse effect changes with elevation through water vapor feedback. An increase of greenhouse gases in the atmosphere due to human activity might result in an increase of the vertical gradient of the greenhouse effect in particular due to water vapor feedback.

The previous section demonstrated that MODTRAN has the ability to calculate clear-sky correlations between temperature, precipitable water and longwave downward radiation correctly. In other words, MODTRAN seems to model the physics of the atmosphere correctly under clear-sky conditions. It can therefore be assumed that MODTRAN is also able to model the relation between the greenhouse radiative flux \(G\) and the longwave downward flux \(LWL\) correctly.

The greenhouse radiative flux was obtained from the model output based on Equation 8.1 by subtracting longwave radiation escaping towards space from estimates of the radiation emitted by the surface. Figure 8.7 (left graph) reveals that \(LWL\) increases significantly with increasing \(G\). A 1 W m\(^{-2}\) increase of \(G\) yields a 2 W m\(^{-2}\) increase in \(LWL\). This doubling is slightly more pronounced at higher altitudes, where the gradient becomes steeper. Even when the two fluxes are normalized by the surface emission (right graph), the increase of \(nLWL\) with \(nG\) and the altitude dependence is still evident. It could therefore be considered, that an increase in \(G\) can best be detected at high mountain stations. This would be true for a constant increase of \(G\) at all altitudes, which is not the case as demonstrated in Table 8.1. Figure 8.8 and Figure 8.9.

![Figure 8.7: MODTRAN calculation of clear-sky greenhouse radiative flux versus longwave downward radiation for four different atmospheres and altitudes. The normalization on the right hand graph was performed by surface emission.](image)

The greenhouse radiative flux has been measured globally by satellite within the Earth Radiation Budget Experiment (ERBE) over the oceans. The emission of the Earth's surface was calculated from sea surface temperatures (also measured by satellite), which have small daily variations and small uncertainties in the emittance. Raval and Ramanathan (1989) calculated correlations between precipitable water, sea surface temperature and the greenhouse radiative flux, which can be compared to our results calculated by the MODTRAN model.

Figure 8.8 shows the normalized clear-sky greenhouse \((nG)\) as a function of surface temperature. MODTRAN calculations of four different atmospheres and altitudes are compared to ERBE-observation of April 1985. \(nG\) increases significantly with surface temperature, even when it is normalized by...
the surface emission. The increase is larger at lower altitudes due to higher temperatures. The slightly higher rate of the sea-surface based ERBE observed increase of \( \frac{dT}{dG} = 3.42 \times 10^{-3} \) in comparison to the model values at 500 m a.s.l. \( \frac{dT}{dG} = 3.31 \times 10^{-3} \) corresponds almost perfectly to the extrapolated sea-level values.

**Figure 8.8:** Temperature normalized clear-sky greenhouse effect as a function of surface temperature. Model values of four different atmospheres and altitudes are compared to ERBE observed values from April 1985.

The variation of the water vapor in the atmosphere is determined largely by the increase in the saturation vapor pressure with temperature, as shown by the model calculations \( \frac{d\ln(PW)}{dT} = 5.65 \times 10^{-2} \), Figure 8.5), the ASRB-data \( \frac{d\ln(PW)}{dT} = 5.14 \times 10^{-2} \), Figure 8.3) and the ERBE-observations \( \frac{d\ln(PW)}{dT} = 5.53 \times 10^{-2} \), not shown). If the observed variation of water vapor causes the G-T coupling revealed by the model and the data (Figure 8.8), the atmospheric absorptivity must scale logarithmically with the precipitable water.

**Figure 8.9:** Normalized clear-sky greenhouse effect as a function of precipitable water. Model values of four different atmospheres and altitude are compared to ERBE observed values from April 1985.
This strong correlation of the greenhouse effect with the precipitable water (PW) is shown in Figure 8.9. A larger increase of nG with water vapor is observed at lower stations due to higher temperatures. The rate of the modelled increase of nG at the lowest altitude (dnG/dPW = 5.84*10^{-2}) in comparison to the ERBE observation (dnG/dPW = 5.76*10^{-2}) is therefore a little bit too high. Nevertheless, our model results of surface temperature, precipitable water and greenhouse effect have been confirmed by the ERBE-observations and thus give compelling evidence that the water vapor feedback is correctly modelled within MODTRAN.

### 8.7 Greenhouse Effect in the Alps

The good agreement between model calculations and measurements justify the investigation of the greenhouse effect for the ASRB-stations. Due to the good correlation between LW↓ and G for all points (Figure 8.7, R²=0.98), the radiative greenhouse flux for the ASRB-stations can be calculated in the same way by transforming the regression equation of all model points (all atmospheres and altitudes) in Figure 8.7 to:

\[
G = -24.37 + 0.527LW↓ \quad (8.3)
\]

This equation yields the clear-sky greenhouse effect (G or \(G_{\text{clear-sky}}\)). The addition of the longwave cloud forcing (Figure 7.5) reveals the all-sky greenhouse effect (\(G_{\text{all-sky}}\), Equation 8.4), which is listed in Table 8.3 for all ASRB-stations.

\[
G_{\text{all-sky}} = G_{\text{clear-sky}} + CF_{LW} \quad (8.4)
\]

**Figure 8.10: Annual and seasonal mean all-sky greenhouse radiative fluxes at all ASRB-stations.**

The clear-sky greenhouse effect shows a clear altitude dependence (annual mean gradient -1.7 W m⁻²/100 m). This is caused by the stronger decrease of LW↓_{surface} in comparison to that of LW↓_{Top}. This effect is reduced by the presence of clouds resulting in a lower altitude gradient (-1.1 W m⁻²/100 m) for the all-sky greenhouse effect (Figure 8.10). The all-sky summer gradient is perturbed by frequent convective clouds at mountains stations below 3000 m a.s.l. During winter, on the contrary, the frequent stratoform cloud layer over the Swiss central plateau causes an unusually large greenhouse effect at Payerne.
TABLE 8.3: Annual and seasonal mean values of the greenhouse radiative flux (clear- and all-sky) and of longwave cloud forcing at all ASRB-stations.

<table>
<thead>
<tr>
<th>[W m⁻²]</th>
<th>Greenhouse effect (clear-sky)</th>
<th>Longwave cloud forcing</th>
<th>Greenhouse effect (all-sky)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Year</td>
<td>Wi</td>
<td>Su</td>
</tr>
<tr>
<td>LOM</td>
<td>121</td>
<td>100</td>
<td>146</td>
</tr>
<tr>
<td>PAY</td>
<td>117</td>
<td>96</td>
<td>142</td>
</tr>
<tr>
<td>DAV</td>
<td>100</td>
<td>82</td>
<td>124</td>
</tr>
<tr>
<td>CIM</td>
<td>99</td>
<td>82</td>
<td>122</td>
</tr>
<tr>
<td>MAE</td>
<td>91</td>
<td>77</td>
<td>110</td>
</tr>
<tr>
<td>VSF</td>
<td>82</td>
<td>67</td>
<td>102</td>
</tr>
<tr>
<td>WFJ</td>
<td>82</td>
<td>66</td>
<td>101</td>
</tr>
<tr>
<td>EGH</td>
<td>77</td>
<td>64</td>
<td>95</td>
</tr>
<tr>
<td>DIA</td>
<td>73</td>
<td>61</td>
<td>93</td>
</tr>
<tr>
<td>GOR</td>
<td>72</td>
<td>59</td>
<td>90</td>
</tr>
<tr>
<td>JFJ</td>
<td>66</td>
<td>56</td>
<td>80</td>
</tr>
</tbody>
</table>

FIGURE 8.11: Annual mean values of longwave downward radiation as a function of the all-sky greenhouse effect for all ASRB-stations. The all-sky greenhouse effect was calculated from the sum of the clear-sky greenhouse effect and the longwave cloud forcing.

The determined all-sky greenhouse effect demands for a correlation with the measured all-sky longwave downward radiation (FIGURE 8.11). The resulting good coefficient of determination ($R^2=0.95$) allows to transform the regression equation to approximate the effect of a possible change in the all-sky greenhouse flux to longwave downward radiation.

\[
G_{all-sky} = 33.10 + 0.40 LW_{all-sky} \tag{8.5}
\]
The rate of all-sky increase of $LW_\downarrow$ with $G$ is even larger ($dLW_\downarrow/dG = 2.50$, Figure 8.11) than for clear-sky conditions ($dLW_\downarrow/dG = 1.94$, Figure 8.7). This implies that a change in the all-sky greenhouse flux is amplified by a factor of 2.5 for the longwave downward radiation. It is thus confirmed that not only the clear-sky $LW_\downarrow$, but also the all-sky $LW_\downarrow$ is a reliable parameter to monitor and detect possible changes in a changing climate.

An attempt to collocate ERBE and ASRB determined clear-sky greenhouse effects revealed similar results. The mid-European ERBE-value from April 1985 was compared with the monthly mean value of April from the lowest ASRB station (Locarno-Monti, most similar to sea surface temperatures). The clear-sky $LW_\downarrow$ of Locarno-Monti for April (267 W m$^{-2}$) yields a $G$ of 114 W m$^{-2}$. This is within the ERBE measured values of $G$ (for the sea at the latitude of mid-Europe), which varies between 100 and 120 W m$^{-2}$.

**Figure 8.12:** Installation of the downward looking instruments at the ASRB-station Weissfluhjoch. The special design of the ASRB-instruments allows to measure the short- and longwave fluxes from the ground and the sky even under harsh alpine conditions.
9 Conclusions

9.1 Synthesis

This study focused on a better understanding and knowledge of the surface radiation fluxes, the cloud forcing and the greenhouse effect in the Swiss Alps. Special emphasis was placed on the measurement method and analysis of longwave radiation. The most important features and findings are summarized in the following.

Longwave Radiation Measurement

Pyrgeometers are best used in shaded mode to accurately measure longwave atmospheric radiation. Direct solar radiation on unshaded pyrgeometers produce an error of 5 to 20 W m\(^{-2}\). If permanent shading is impossible this error can be successfully corrected with the help of a fixed shadow band. With an ASRB-like instrumentation and modified pyrgeometer absolute measurements are within 1%. The correction algorithm is simple (based on global radiation), seasonally independent and may even be improved with the help of additional measurements.

Clear-Sky Detection

The distinction between clear- and all-sky situations is a necessary condition for many applications in climate research. Longwave downward radiation is an excellent element to continuously monitor the sky. Together with the station temperature and humidity, longwave downward radiation was used to develop an algorithm to determine clear-sky. This algorithm provides a clear-sky index, which allows to detect clear-sky conditions independent of day-time and cloud observations (CHAPT. 5). If refined, in this method may even be the potential to distinguish between different cloud amounts.

Surface Radiation Fluxes

Annual and seasonal mean values of short- and longwave radiation fluxes from the surface and the ground were determined for each ASRB-station. Besides of some regional differences the altitude dependence of the different fluxes was clearly revealed (CHAPT. 6).

The annual mean values of global radiation demonstrate an altitude gradient of 1.3 W m\(^{-2}\)/100 m. Compared to the north, the stations on the south of the Alps show slightly larger absolute values. Shortwave net radiation is almost constant around 80 W m\(^{-2}\) at all mountain stations. The lowland stations in contrast show a higher value of about 120 W m\(^{-2}\) due to a lower albedo.

The annual mean values of longwave atmospheric radiation yield an altitude gradient of -2.9 W m\(^{-2}\)/100 m. Again compared to the north the stations on the south side of the Alps show slightly lower absolute values. Longwave net radiation is always negative (between -50 and -150 W m\(^{-2}\)) and demonstrates neither pronounced seasonal differences nor a clear altitude gradient. The reason is the cloud amount, which influences longwave net radiation the most.

Annual mean net radiation is positive at all stations due to the overbalance of the shortwave net radiation. The absolute values vary from 50 W m\(^{-2}\) at the lowest stations to almost zero at the highest mountain stations. The available energy for the non-radiative components of the surface energy balance (latent and sensible heat) is therefore very low at mountain stations. Hence, the annual mean
values reveal an altitude gradient of -1.5 W m\(^{-2}\)/100 m. The summer mean values show similar net fluxes around 120 W m\(^{-2}\) at all altitudes. The winter mean values are all negative, except at Payerne (+3 W m\(^{-2}\)) as a result of a frequent stratoform cloud layer.

**Cloud Forcing**

The development of the Clear-Sky Index allowed to determine the impact of clouds over the Swiss Alps (CHAP. 7). Shortwave cloud forcing is negative at all stations with lower values at high altitudes due to higher surface albedo. The high albedo of snow is also responsible for the low winter values (about -5 W m\(^{-2}\)) at the mountain stations. The annual mean values reveal an altitude gradient of shortwave cloud forcing of 0.51 W m\(^{-2}\)/100 m.

Longwave cloud forcing is always positive with lower values at the lower stations, which is a consequence of larger distances to the cloud base. The annual mean values reveal an altitude gradient of 0.58 W m\(^{-2}\)/100 m. The highest absolute values of short- and longwave cloud forcing were observed in summer at mountain stations due to convective clouds.

The annual mean values of net cloud forcing revealed to be almost zero at the lower stations due to cancellation of short- and longwave cloud forcing. Stations above 2000 m a.s.l. yield positive annual values. Clouds therefore seem to have little effect on average at low altitudes, but a warming effect at high altitudes. The altitude gradient of the annual mean values is calculated as 1.1 W m\(^{-2}\)/100 m. Summer values are always negative and winter values positive, which is predominately an effect of the seasonal change of the shortwave cloud forcing.

**Greenhouse Effect**

The greenhouse radiative flux can not be directly measured, but longwave downward radiation is directly related to the greenhouse effect, as shown in SECTION 8.6. This dependence and the small year to year variation (1% standard deviation) makes longwave downward radiation one of the most promising elements to monitor climate change. Theoretical considerations and model calculations reveal that it may be particularly worthwhile to monitor longwave downward radiation at high altitudes. A radiative transfer model allows to investigate the relation between the greenhouse flux and the longwave flux. The good correlation made it possible to calculate the greenhouse flux at all ASRB-stations. Lower stations yield a higher greenhouse flux due to higher temperature and more water vapor. The correlation also showed that a change in the greenhouse effect amplifies longwave downward radiation by a factor of 2.5.

**9.2 Achievements**

All the ASRB-stations have been successfully operational for four years or more. The special design of the ASRB-instrumentation guarantees high quality measurements even under harsh alpine conditions. The high resolution data are controlled and corrected daily and filled in a database once a month. A new algorithm has been developed to automatically detect clear-sky conditions with the help of longwave downward radiation. The results presented in this study demonstrate the valuable and many-sided information content of the ASRB radiation network.

For the first time the altitude dependence of the different radiation fluxes have been measured with high accuracy over a longer time span. Also for the first time seasonal and annual mean values of all radiation fluxes (distinguished between clear-and all-sky) have been determined at altitudes up to 3500 m a.s.l.. The impact of clouds has been revealed by the calculation of the seasonal mean values of the cloud forcing for all stations. These quantities were set in relation to the current greenhouse
effect, which was calculated for the first time over Switzerland. The impact of the expected warming on different altitudes was computed and discussed with respect to possible changes in the longwave downward radiation.

The ASRB-network has delivered high altitude radiation data, which are not only necessary for a better understanding of the earth-atmosphere interaction in a mountain range, they are also used by different research groups to test or improve energy balance models, in particular with regard to parameterization of short- and longwave radiation in mountain environments.

9.3 Outlook

The Alps being a unique place to investigate altitude dependent atmospheric processes may also play an important role in monitoring the reaction of the radiation fluxes due to climate change. Some climate models predict greenhouse warming for midlatitude to be larger in the troposphere than at the surface (Hansen et al. 1988). Hence, the best place to observe greenhouse warming may be at high mountain stations. It would therefore be very valuable if a new ASRB-station could be placed somewhere above 4000 m a.s.l. It may, on the other hand, be questioned whether the three stations Eggishorn, Diablerets and Gornegrat, which are at about the same altitude between 2800 and 3100 m a.s.l., will all be needed in the future. In any case, it is better to thoroughly maintain a few representative, high quality radiation stations rather than collect as much data as possible. Wielicki et al. (1995) emphasized this fact by writing the following sentence about the possibility of directly monitoring climate change with the help of radiation measurements: “One of the most difficult aspects of this challenge will be to keep the long-term perspective needed to obtain consistent climate records of cloud and radiation budget parameters.”

This work introduces many radiation issues - some problems could be discussed and others still have to be interpreted. There are many scientific questions to answer, which have not been addressed, if the ASRB-network also continues to deliver high quality radiation data. In a few years it should be possible to create a characteristic radiation and cloud climatology. Proper statistics with maximum and minimum mean values will help detect future trends. A longer time series will also help investigate the observed radiation differences between the northern and the southern side of the Alps. The radiation climatology makes it possible to compare and adjust GCM-calculated values and gradients with ASRB mean data. A characteristic cloud climatology will allow to quantify the differences between overcast-sky (not all-sky) and clear-sky, which has not yet been done.

The representative albedo of a alpine region is a question which has not been answered yet - although this element is crucial for the accurate determination of the radiation budget and the cloud forcing.

An unsolved problem regarding the clear-sky index is the special case of the cirrus clouds. They have not been especially considered, although there are quite frequent and sometimes hardly to detect by synoptical observation.

New methods to determine the water vapor content with Precision Filter Radiometer (PFR) or Global Positioning System (GPS) instruments, will allow to investigate the relation between longwave downward radiation and the measured water vapor content instead of the station humidity. Future satellite experiments over Europe (Wielicki et al. 1995) will allow to collocate the surface measurements with top of atmosphere data. This will make it possible to directly measure the greenhouse effect over the Alps. It will also reveal the differences between the surface and the top of atmosphere cloud forcing.

The ASRB-stations therefore build a unique network to study radiation processes at different altitudes and to monitor the change in longwave radiation due to increasing greenhouse gases.
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Appendix

A Horizon at ASRB-Station

Figure A-1: Horizon at the ASRB-station Locarno-Monti.

Figure A-2: Horizon at the ASRB-station Payerne.
**Figure A-3:** Horizon at the ASRB-station Davos.

**Figure A-4:** Horizon at the ASRB-station Cimetta.
FIGURE A-5: Horizon at the ASRB-station Männlichen.

FIGURE A-6: Horizon at the ASRB-station Weissfluhjoch.
FIGURE A-7: Horizon at the ASRB-station Eggishorn.

FIGURE A-8: Horizon at the ASRB-station Diablerets.
FIGURE A-9: Horizon at the ASRB-station Gommergrat.

FIGURE A-10: Horizon at the ASRB-station Jungfraujoch.
B Seasonal Cycle of the Radiation Fluxes

The following figures show the seasonal cycle of all measured radiation fluxes at each ASRB-stations. A daily mean value (dot) for each day was calculated based on the period from 1995-1998. This data-set was smoothed by a 60 day running mean (solid line).

The expanded stations Payerne (inclusion of BSRN-data), Davos, SLF Versuchsfeld and Weissfluhjoch measure down- and upward fluxes. This allowed us to calculate the net fluxes at these stations, which are also shown in this appendix. Together with the information about the cloud cover, which was taken from the daily mean clear-sky index, it was also possible to calculate the short- and long-wave cloud forcing. Besides these radiation fluxes, air temperature and humidity are also plotted in the following figures for better understanding.

The other stations do not measure the upward fluxes. It was therefore not possible to calculate the daily mean values of albedo shortwave cloud forcing. Longwave upward radiation, however, was calculated from the air temperature with an emissivity of 1. The seasonal cycle of longwave upward radiation, longwave net radiation and longwave cloud forcing has therefore to be treated with caution at these stations.

Short- and longwave downward clear-sky fluxes (dashed lines) are also plotted for comparison. The clear-sky cycles were found by averaging all clear-sky moments of a day during the four investigated years. These daily mean values were also smoothed by a 60 day running mean (cf. CHAPT. 6).
Figure B-1: Seasonal variation (60 day running mean=solid lines) of the measured daily mean (dots) radiation fluxes at the ASRB-station Payerne.
FIGURE B-2: Seasonal variation (60 day running mean=solid lines) of the measured daily mean (dots) radiation fluxes at the ASRB-station Davos.
Figure B-3: Seasonal variation (60 day running mean=solid lines) of the measured daily mean (dots) radiation fluxes at the ASRB-station SLF Versachsfeld.
FIGURE B-4: Seasonal variation (60 day running mean=solid lines) of the measured daily mean (dots) radiation fluxes at the ASRB-station Weissfluhjoch.
Figure B-5: Seasonal variation (60 day running mean=solid lines) of the measured daily mean (dots) radiation fluxes at the ASRB-station Locarno-Monti.
FIGURE B-6: Seasonal variation (60 day running mean=solid lines) of the measured daily mean (dots) radiation fluxes at the ASRB-station Cimetta.
FIGURE B-7: Seasonal variation (60 day running mean=solid lines) of the measured daily mean (dots) radiation fluxes at the ASRB-station Männlichen.
FIGURE B-8. Seasonal variation (60 day running mean=solid lines) of the measured daily mean (dots) radiation fluxes at the ASRB-station Eggishorn.
FIGURE B-9: Seasonal variation (60 day running mean=solid lines) of the measured daily mean (dots) radiation fluxes at the ASRB-station Diablerets.
Figure B-10: Seasonal variation (60 day running mean=solid lines) of the measured daily mean (dots) radiation fluxes at the ASRB-station Gomergrat.
FIGURE B-11: Seasonal variation (60 day running mean=solid lines) of the measured daily mean (dots) radiation fluxes at the ASRB-station Jungfraujoch.
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Curriculum Vitae

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Education

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