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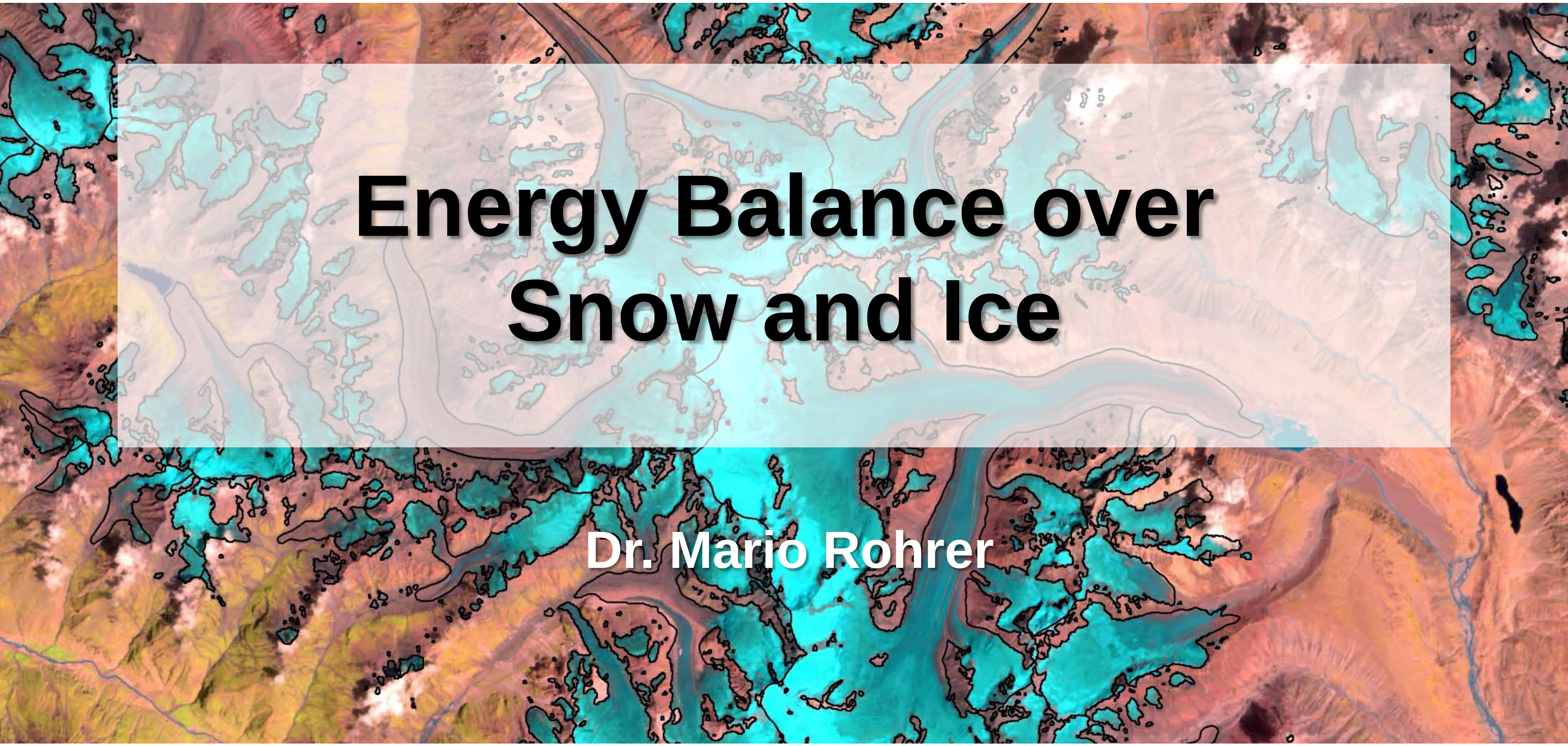


Energy Balance over Snow and Ice

Dr. Mario Rohrer

IHCAP – Indian Himalayas Climate Change Adaptation Programme
Capacity building programme “Cryosphere” Level-2 (January, 5–February 13, 2015)

Contents

2. Blackbody radiation

Radiation emitted by bodies

- **All objects** having a temperature > 0 K **emit radiation**
- **Sun** and **earth's surface** behave approximately as **black bodies**.
- ***Blackbody*** = any object that is a perfect emitter and a perfect absorber of radiation:
  ⇨ all incident radiation is completely absorbed.
  ⇨ maximum possible emission is realized.

Stefan-Boltzmann Law

- The **Stefan-Boltzmann law** relates the total amount of radiation emitted by an object to its temperature:

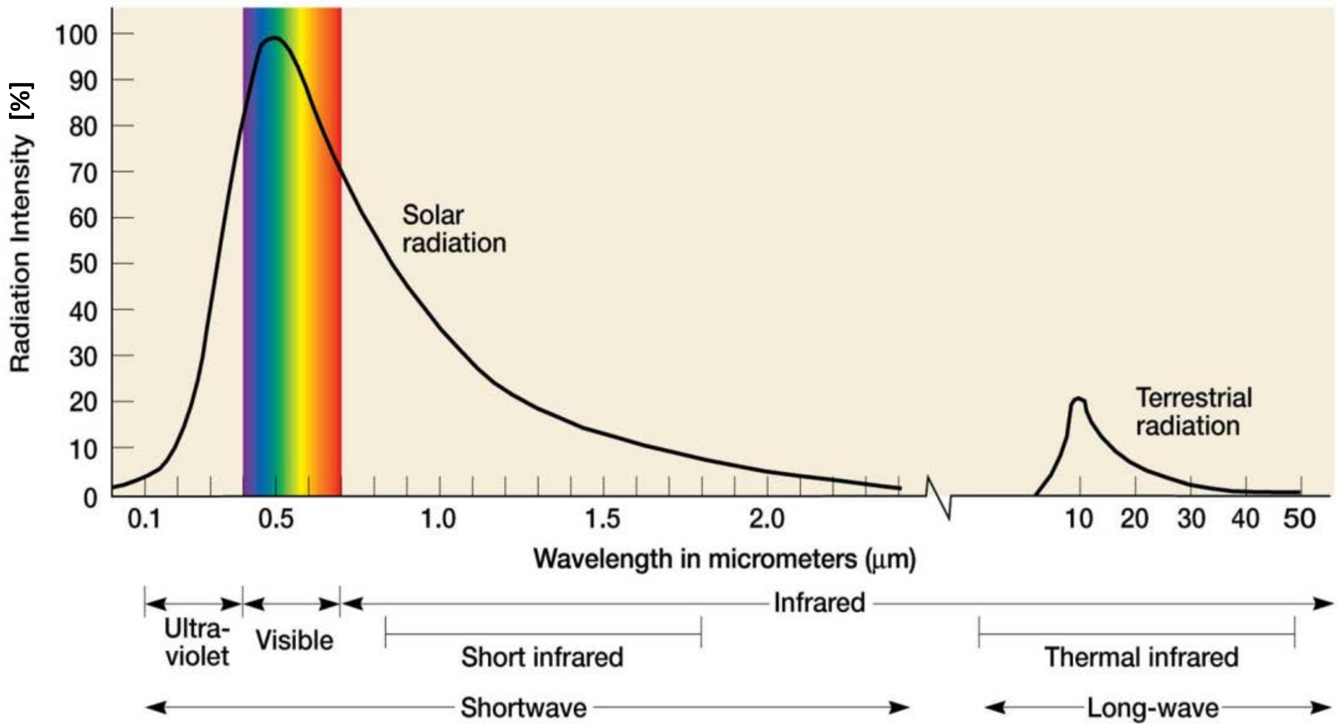
$$\boxed{E = \sigma T^4}$$

E = total amount of radiation emitted by an object [W m^{-2}]
σ = Stefan-Boltzmann constant = 5.67 x 10^{-8} [W m^{-2} K^{-4}]
T = temperature of the object [K]

2. Blackbody radiation

Emission spectrum of the Sun and Earth

- The electromagnetic radiation of the Sun is nearly that of a black body at about 5500K.
- The main range of solar radiation includes ultraviolet radiation (UV, 0.001-0.4 µm), visible radiation (light, 0.4-0.7 µm), and infrared radiation (IR, 0.7-100 µm).
- The earth is much cooler than the sun and emits infrared radiation.



2. Blackbody radiation

Peak of blackbody radiation (Wien's displacment law)

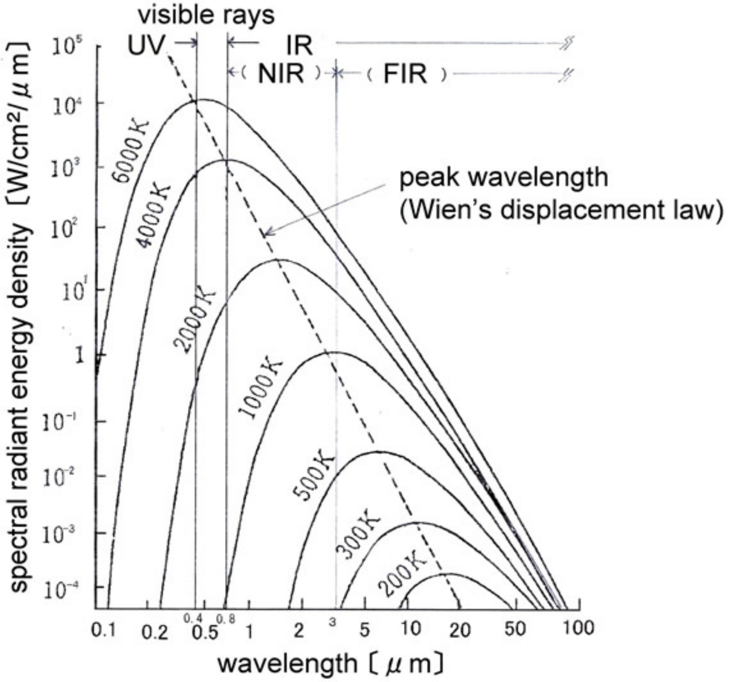
- All **blackbodies emit radiation** and the **wavelength** and **energy characteristics** (**spectrum**) of that radiation are **determined** solely by the **blackbody's temperature**.
- **When** the **temperature** of a **blackbody radiator increases**, the overall radiated **energy increases** and the **peak of the radiation curve moves to shorter wavelengths**.

This relationship is called:

Wien's Displacement Law

$$\boxed{\lambda_{max} = 2.89 * 10^{-3} / T}$$

λ_{max}: wavelength of peak emission [m]
T: temperature [K]



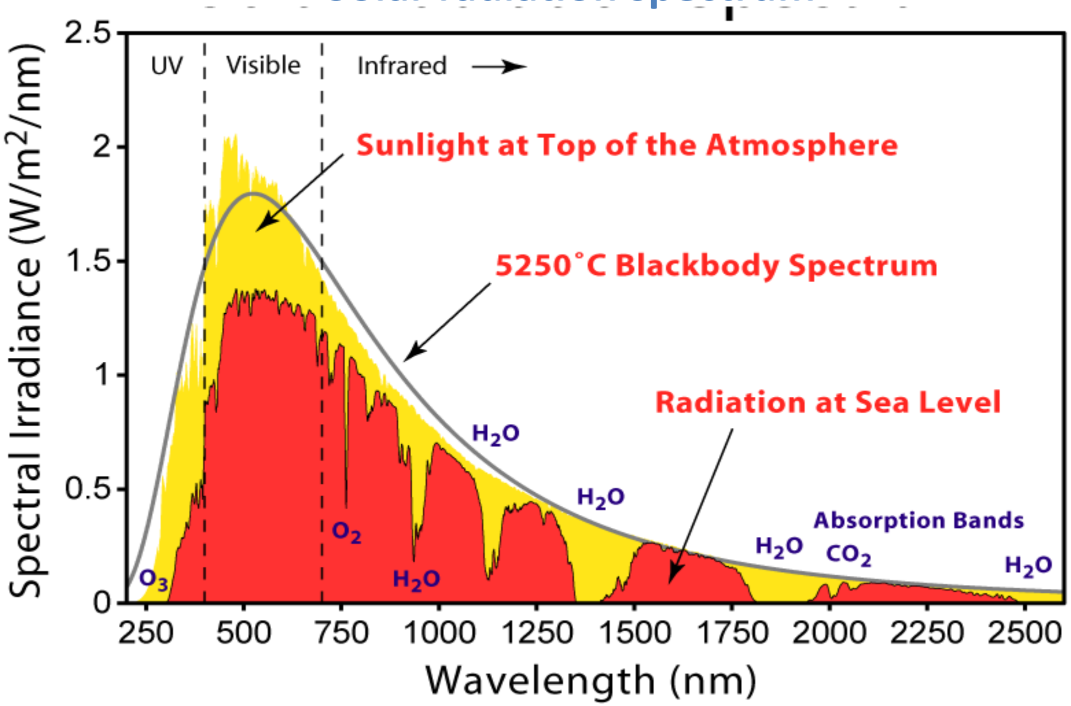
Contents

3. Solar radiation

Solar radiation

- The energy flow within the **sun** results in a **surface temperature** of around **5800 K**.
- The spectrum of the **radiation** from the **sun** is similar to that of a **5800 K blackbody**.
- The solar constant is defined as the **flux of solar radiation** at the **top of the atmosphere (TOA)** at the **mean distance** between the **Earth and the Sun**.

Solar radiation spectrum



http://en.wikipedia.org/wiki/Image:Solar_Spectrum.png

- It is considered to be **constant for many practical purposes**: 1366 W m^{-2}.
- The **average incoming solar radiation** is **one fourth** the solar constant or ~342 W m^{-2}.

3. Solar radiation

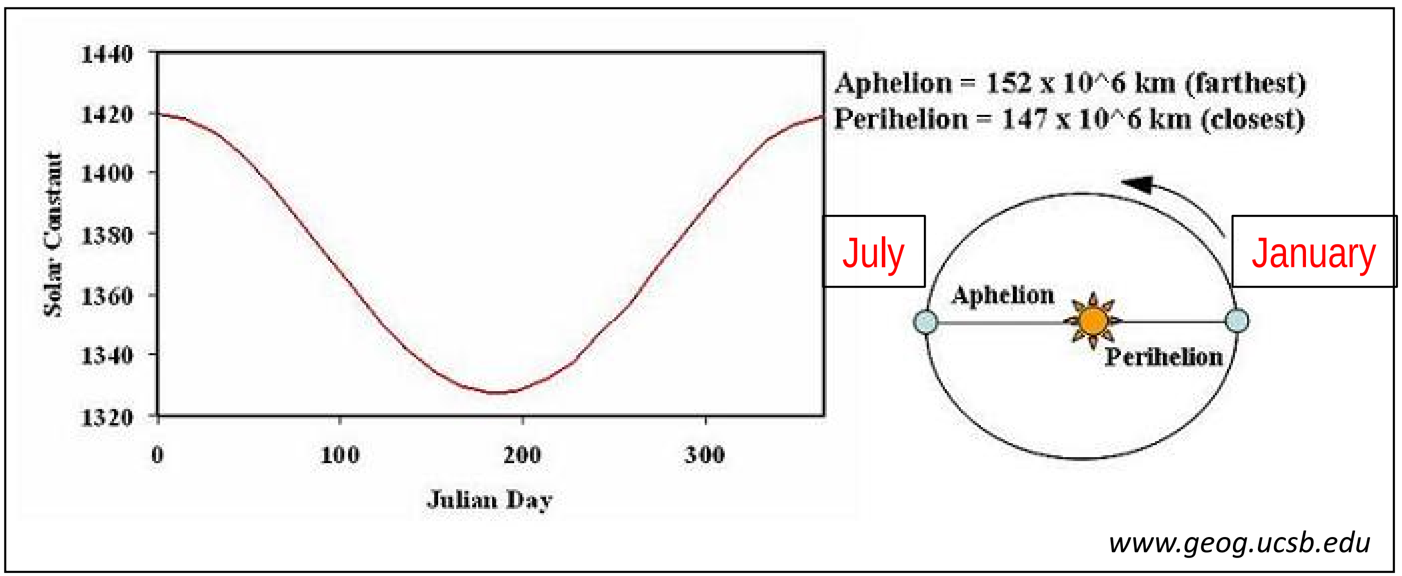
Variation of solar radiation

Interannual variation

- The **actual direct solar irradiance** at the **top of the atmosphere** fluctuates by about **6.9%** during a year due to the Earth's varying distance from the Sun.

- **Radiation intensity** is **proportional** to **the square inverse of the sun-earth distance**.

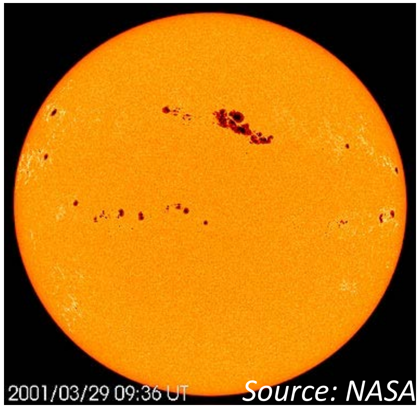
- The **minimum** of **solar constant** is **in early July**, the **maximum** in **early January**.



3. Solar radiation

Variation of solar radiation

Sunspots



Source: NASA

- Sunspots are regions on the solar surface that appear dark because they are cooler than the surrounding photosphere.
- The temperature of sunspots is about ***4000 – 4500 K***.
- Sunspots are areas where the ***magnetic field*** is about 2,500 times stronger than Earth's, ***much higher*** than anywhere else on the Sun.
- These spots are ***much bigger than the Earth***; they can be over 10 times the diameter of the Earths.
- Sunspots move across the surface of the Sun.
- They only last for one to two weeks, but the ***number of sunspots*** follows an 11-year cycle.

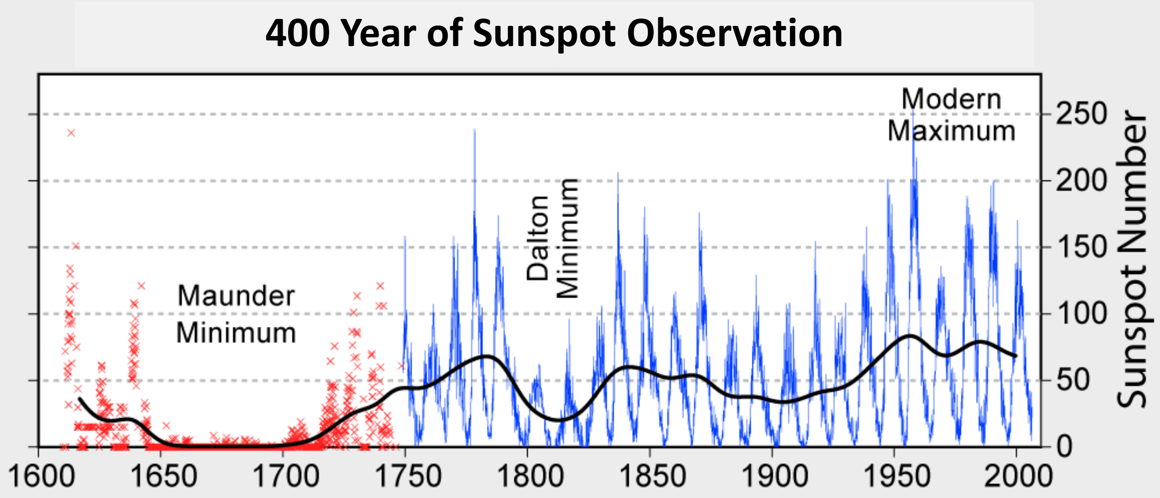


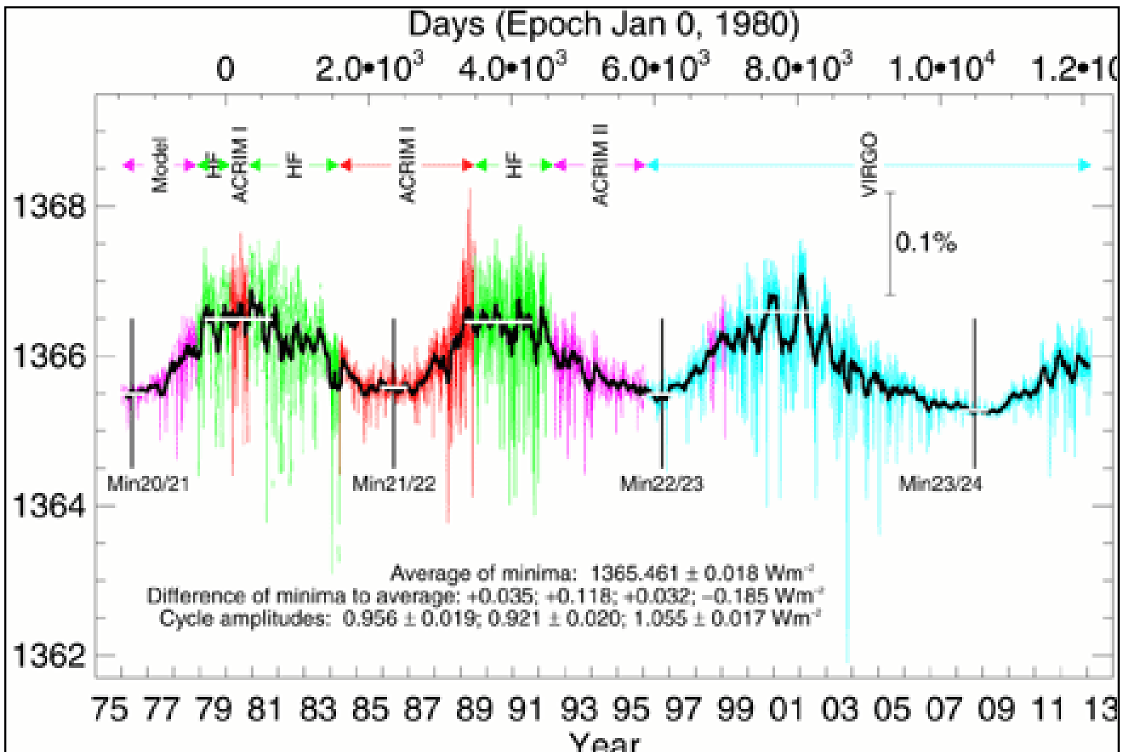
Image : Robert A. Rohde

3. Solar radiation

Variation of solar radiation

11-years variability of the solar constant

- Variation in solar radiation was too small to detect with technology available before the satellite era.
- Since 1979 ***satellite measurements*** of absolute radiative flux became available.



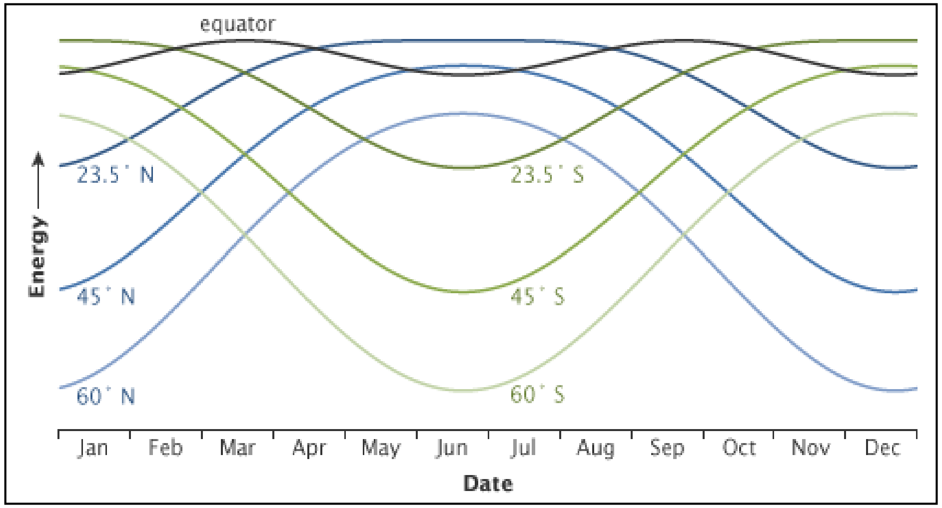
World Radiation Center, Davos

composite of total solar irradiance

- Periods of ***high solar activity*** correspond generally to ***sunspot maxima***.
- The surrounding margins of sunspots are hotter than the average. Overall, more sunspots increase the Sun's solar constant.
- The ***variation*** caused by the sunspot cycle to solar output is relatively small: ~ 1.3 W/m².

Incident solar radiation

- **Solar illumination varies** in **space** and **time**.
- The **annual amount of incoming solar energy** varies considerably from **tropical latitudes** to **polar latitudes**.
- It also **varies considerably** from **season to season** (middle and high latitudes).



http://earthobservatory.nasa.gov/Features/EnergyBalance/page3.php

Conservation of energy

As long as the electromagnetic radiation from the Sun travels through empty space it remains intact because of the vacuum.

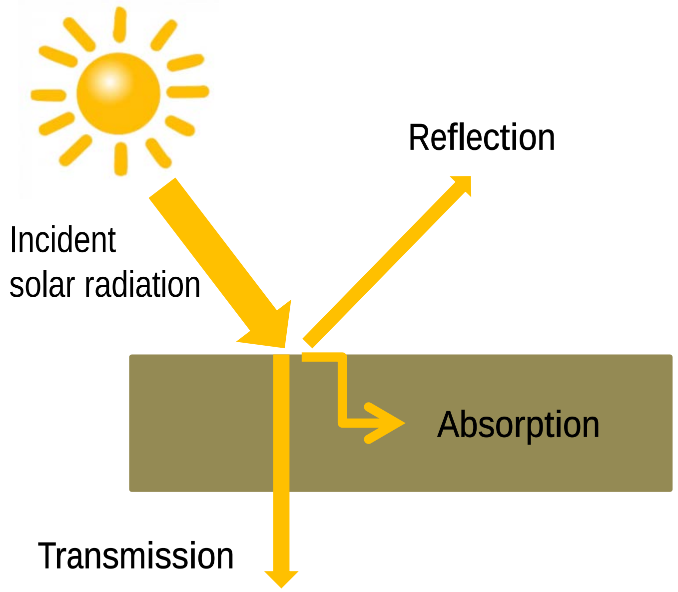
When the **solar radiation** encounters a **parcel of matter** (solid, liquid or gas), it goes through one or all of three processes:

Reflection + **Absorption** + **Transmission** = 100%

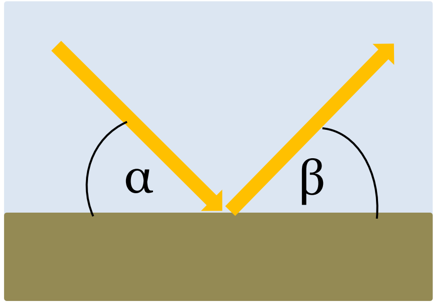
Reflection:
Occurs when radiation is redirected as it strikes a target.

Absorption:
Occurs when radiation is absorbed into the target (heating).

Transmission:
Occurs when radiation passes through a target without significant attenuation.



Planetary Reflection

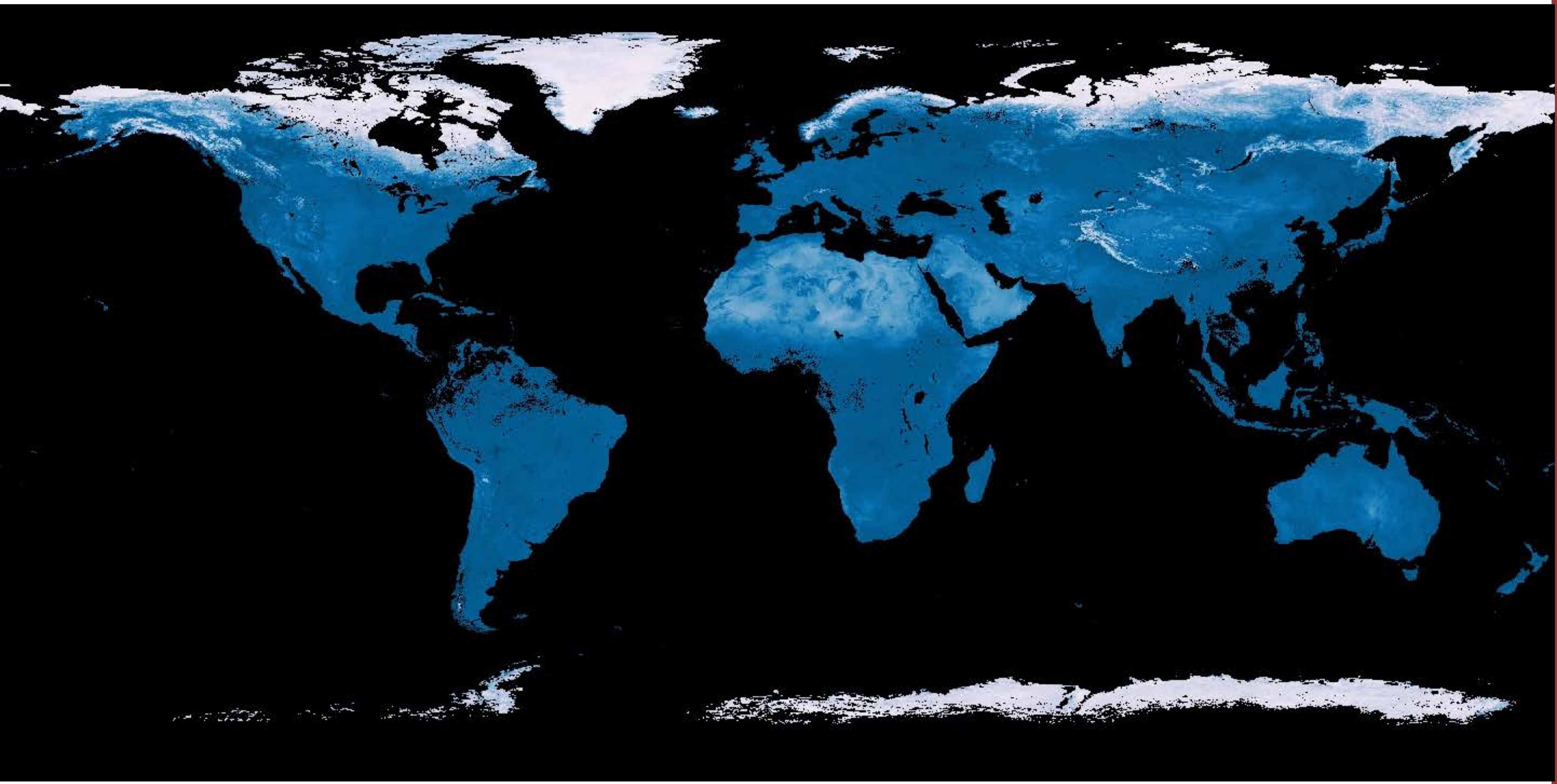


- Reflection is a process where sunlight is redirected after striking an atmospheric particle (clouds, gas molecules, aerosols) or the surface.
- Angle α of incident light = angle β of reflected light.
- As in the atmosphere, some of the radiation received at the Earth's surface is redirected back to space by reflection.
- A considerable **portion of solar radiation** is **reflected** back **into outer space** upon striking the **uppermost layers of the atmosphere**, and also from the **tops of clouds**, when light is intercepted by particles of liquid and frozen water.
- ***Reflectivity*** of the surface is often described by the term surface ***albedo***.
- The **Earth's average albedo**, reflectance from both the atmosphere and the surface, is **about 30%.**
- **Clouds** contribute **about 2/3 of planetary albedo**.

3. Solar radiation

Interaction of solar radiation

Earth's land surface albedo



Albedo of May 2014. (*NASA Earth Observations. http://neo.sci.gsfc.nasa.gov*)

Absorption

- **Absorption** is defined as a **process** in which **solar radiation is retained** by a substance and ***converted into heat energy***. The **creation of heat energy also causes** the substance **to emit its own (longwave) radiation**.

- **When** the **solar radiation reaches the E-A-System** it is **partially absorbed at various levels in the atmosphere and by the Earth surface**.

- **Solar radiation absorption** is **uneven** in **both space and time** and this gives rise to the **complex pattern** and **seasonal variation** of our **climate**.

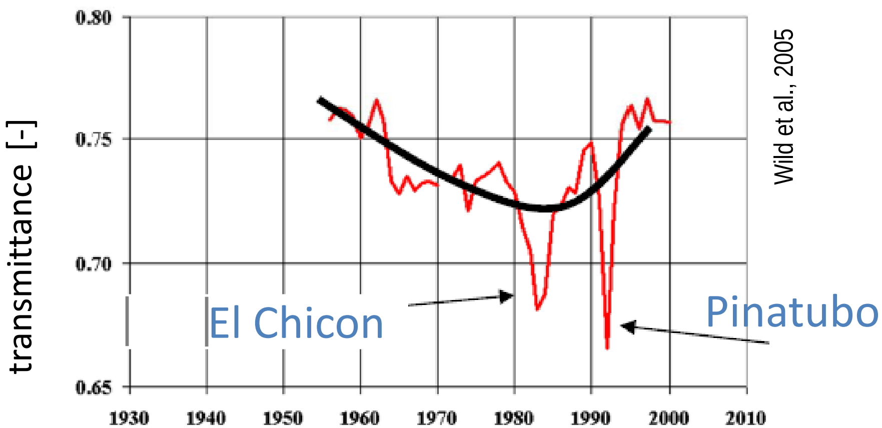
- The **longwave radiation emitted by the Earth's surface** is **mostly absorbed** in the **atmosphere** by **greenhouse gases** and **clouds**.

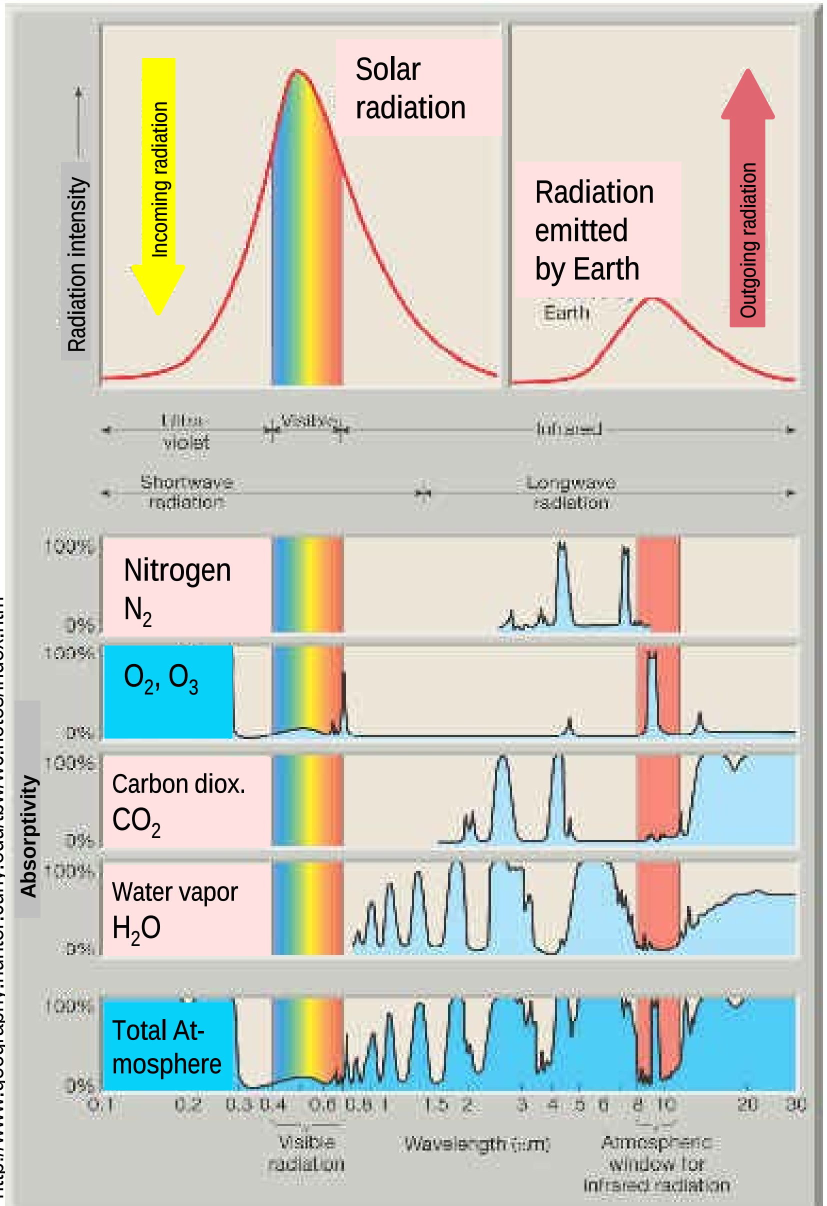
Transmission

- **Transmission: Sunlight passing through atmosphere without being altered**.
- For **certain wavelengths atmosphere is largely transparent**, so called **atmospheric windows** occur.
- The spectral band of **visible light** is one of these **atmospheric windows**.
- Due to transmission
  ⇨ radiation arrives at surface directly from the Sun
  ⇨ the disk of the Sun is visible
  ⇨ patterns of strong light and shadow are created

These **changes in the transmittance** of the cloudless atmosphere, can **partly** be **related** to **changes in the aerosol load due to volcanic eruptions**.

Annual mean total zenith transmittance





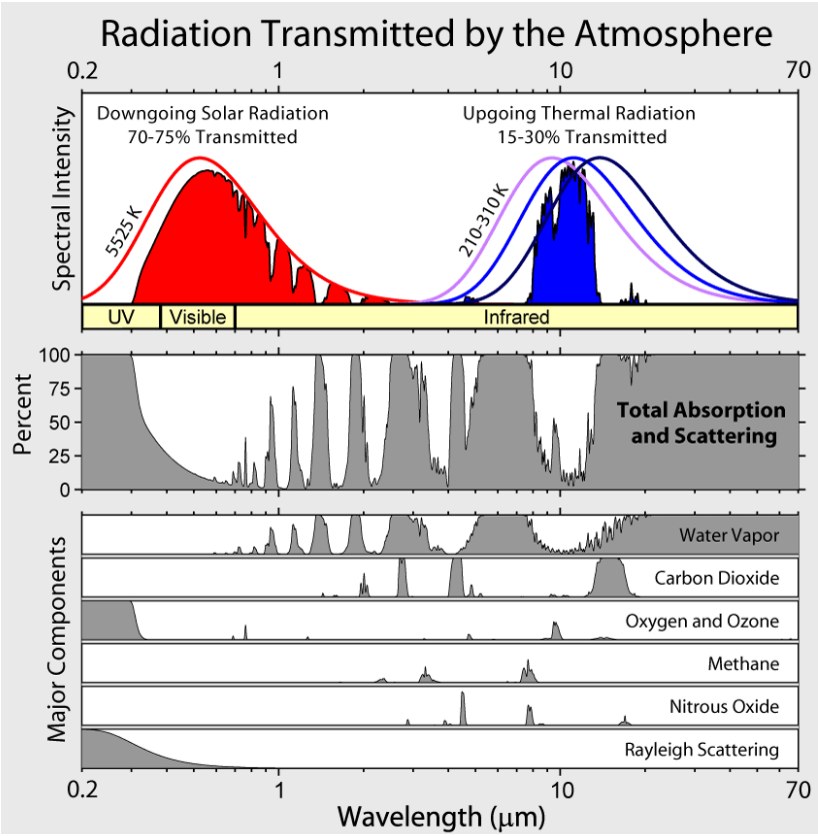
http://www.geography.hunter.cuny.edu/tbw/wc.notes/index.htm

Atmospheric windows

- Some types of electromagnetic radiation easily pass through the atmosphere, while other types do not.
- The **ability of the atmosphere to allow radiation to pass through** it is referred to as its ***transmissivity***, and **varies with the wavelength/type of the radiation and the concentration of gases and aerosols**.
- The **gases that comprise our atmosphere absorb radiation in certain wavelengths** while **allowing radiation with differing wavelengths to pass through**.
- The atmosphere as a whole is quite transparent to solar radiation between ***0.3 and 0.7 μm***, which includes the band of ***visible*** light.
- In the zone between ***8 and 12 μm*** longwave ***infrared*** radiation can pass quite easily.
- Through this atmospheric window most of the earth/atmosphere-system longwave loss to space occurs.

3. Solar radiation

atmospheric window



Atmospheric windows

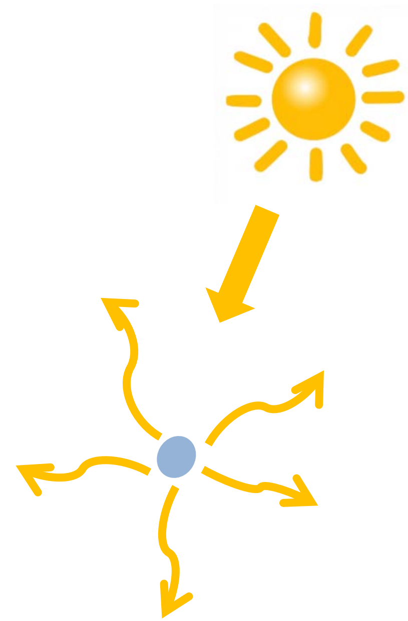
Atmospheric gases also absorb solar energy at certain wavelength intervals called ***absorption bands***, in contrast to the wavelength regions characterized by high transmittance of solar radiation called atmospheric ***transmission bands***, or atmospheric windows.

The degree of absorption of solar radiation passing through the outer atmosphere depends upon the component rays of sunlight and their wavelengths. The gamma rays, X-rays, and ultraviolet radiation less than 200 nm in wavelength are totally absorbed by oxygen and nitrogen.

Most of the radiation with a range of wavelengths from 200 to 300 nm is absorbed by the ozone (O_3) layer in the upper atmosphere.

Scattering

- In addition to reflection, absorption and transmission, scattering is an other way of solar radiation to interact with any object whether in the form of solid, liquid or gas.
- Different from reflection, where radiation is deflected in one direction, some particles and molecules found in the atmosphere have the ability to scatter solar radiation in all directions.
- The scattered radiation is partly returned to space and partly continues its path through the atmosphere.
- Scattering does not change the striking light ray's wavelength.



Scattering

- If scattering did not occur in our atmosphere the daylight sky would be almost black, like the outer space.
- Solar radiation that has been modified by scattering is called diffuse solar radiation.

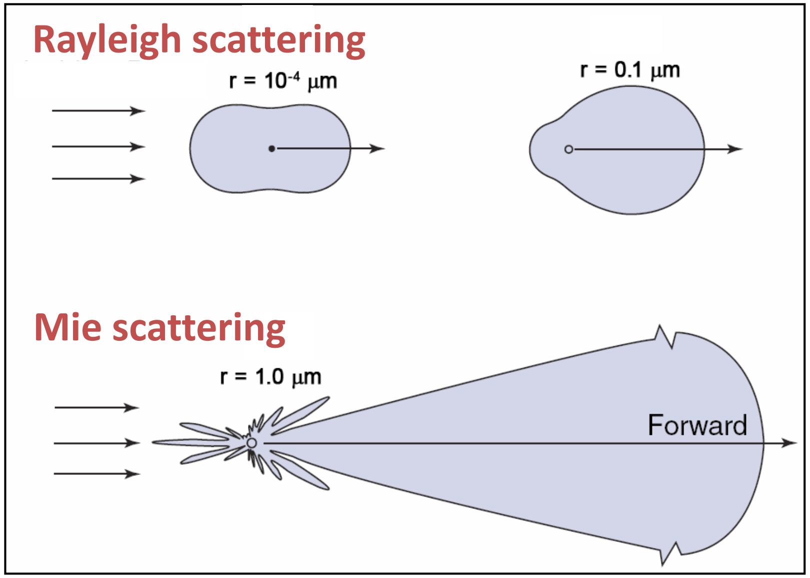
For flat terrain was found that:

- During clear and cloudless days, with low aerosol content, diffuse radiation is about 10 to 14% of the total solar radiation received at the earth's surface.
- Almost only diffuse radiation may reach the earth's surface during extremely cloudy days.

Scattering

- In the atmosphere, the particles responsible for scattering range from gas molecules (~ 10^{-4} µm) to large raindrops and hail particles (~ 10^{4} µm).
- We can broadly distinguish the following categories:

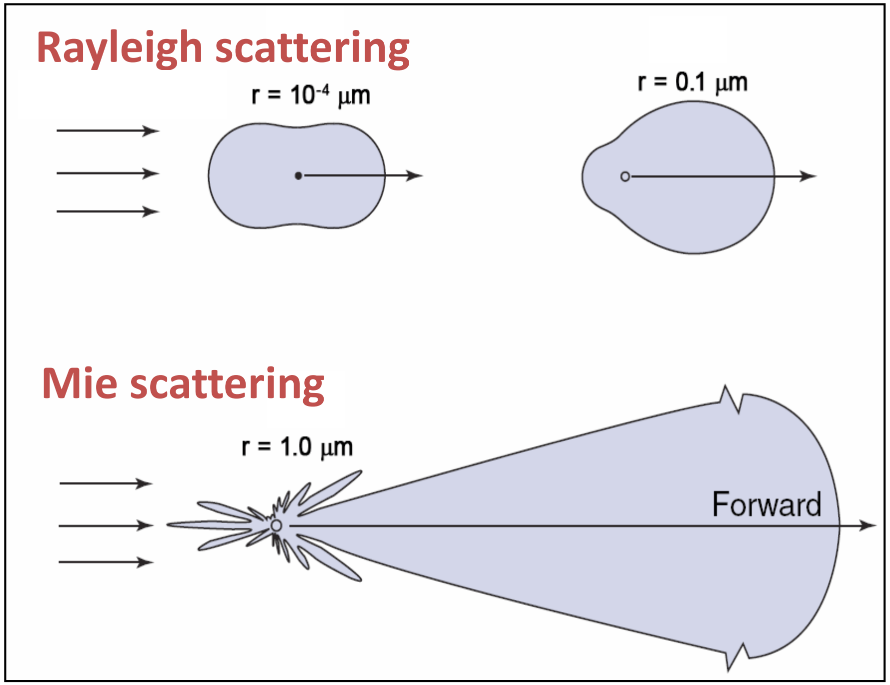
Type of particle	Particle diameter
gas molecules	~ 10^{-4} µm
solid aerosols	0.1 – 1 µm
haze water drops	0.1 – 1 µm
cloud water drops	1 – 10 µm
cloud ice particles	1 – 100 µm
large raindrops, hail	~10^{4} µm



Scattering

- Based on the size of the scattering particles, we distinguish between:

⇨ Rayleigh scattering: particle diameter << wavelength of the incident beam
⇨ Mie scattering: particles diameter >= wavelength of the incident beam



- The scattering from molecules and very tiny particles (< 1 /10 wavelength) is predominantly **Rayleigh scattering**.
- It is characterized by ***symmetry between forward and backward scattering***.
- For particle sizes larger than a wave-length, **Mie scattering** predominates.
- This scattering produces a pattern like an antenna lobe, with a sharper and ***more intense forward lobe*** for larger particles.

Rayleigh Scattering

- Scattering of solar radiation by ***air molecules***
- The particles are much smaller than the wavelength.
- Amount of scattering is ***inversely proportional to the 4th power of the wavelength***.
- Gives the atmosphere its blue colour:
  - ➢ Rayleigh scattering is much greater for blue light than for other colors due to its shorter wavelength.
  - ➢ About 8 times more blue light than red light is scattered.

Paolo Motta, Italy

- Sunsets and sunrises appear reddish because the blue light in the direct light is removed by scattering during the long path through the atmosphere.
  - ➢ Only the longest wavelengths like orange and red of the spectrum remain.

Mie Scattering

- Particles diameter >= wavelength of the incident beam.
- Mie scattering is caused by, ***dust***, ***smoke***, ***water droplets***, ***pollen*** and other particles with particle sizes more than ten times the wavelength of the components of solar radiation.
- Mie scattering occurs mainly ***in the lower portion of the atmosphere***.
- Large particles in the atmosphere are able to scatter all wavelengths of white light equally
- Since the amount of Mie scattering is equal for all wavelengths, clouds and fog appear white/grey although their water particles are colorless.

Spectral bands of incoming solar energy and atmospheric effects

Band	Wavelength [nm]	Atmospheric Effects
Gamma ray	< 0.03	Completely absorbed by the upper atmosphere
X-Ray	0.03 - 3	Completely absorbed by the upper atmosphere
UV (B)	3 - 300	Completely absorbed by oxygen and ozone in the upper atmosphere
UV (A)	300 - 400	Transmitted through the atmosphere, but atmospheric scattering is severe
Visible	400 - 700	Transmitted through the atmosphere, with moderate scattering of the shorter waves
IR	700 - 14000	Absorption at specific wavelengths by carbon dioxide, ozone, and water vapour, with two major atmospheric windows

Global radiation

Upon entering the atmosphere, solar radiation is split into direct and diffuse components due to scattering.

Direct radiation
Radiation that comes directly from the Sun with minimal attenuation by the Earth's atmosphere or other obstacles (*beam solar radiation*).

Diffuse radiation
Solar radiation reaching the Earth's surface that comes from the complete sky hemisphere without direct radiation.

direct + diffuse = global

Global radiation:
The sum of direct and diffuse radiation - received by a horizontal surface from the upper hemisphere – is called global radiation.

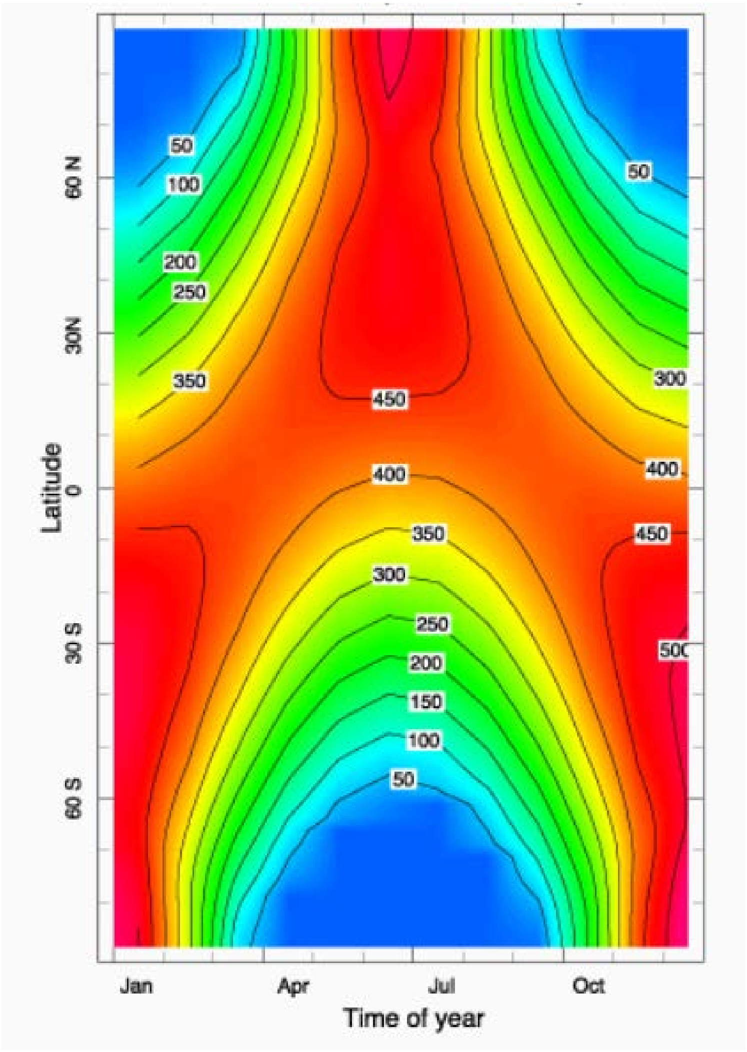
3. Solar radiation

Global radiation

	Influence	Importance
latitude		
altitude		
exposition		
season		
time of day (solar elevation)		
albedo of the ground		
clouds		
humiditiy		
aerosol concentration		
ozon		
water vapour		

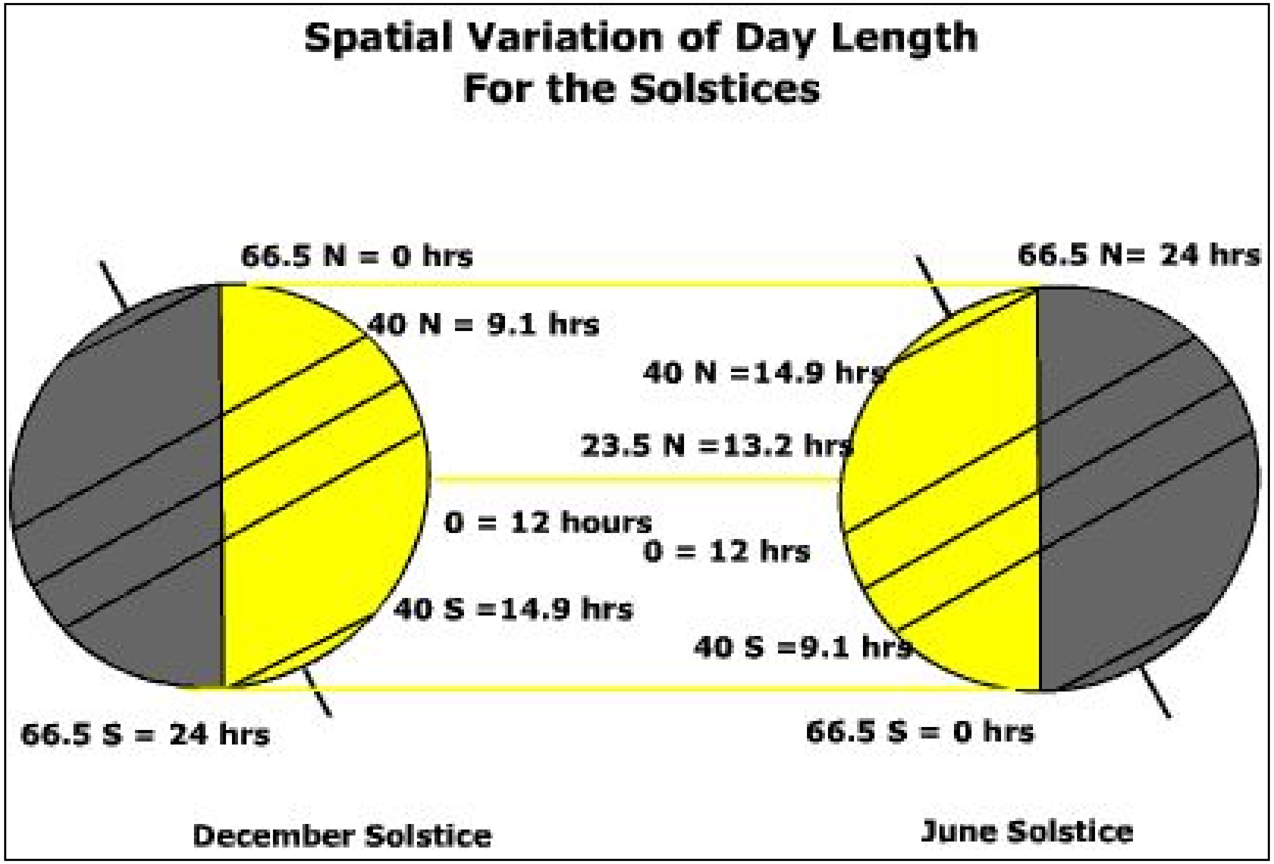
Global radiation

Latitude-Time Distribution of Incoming Solar Radiation at the Top of the Atmosphere [W/m2]



- The figure combines the effects of
  - change in **incidence angle** with latitude
  - **time of year**
  - number of **hours of sunlight** during the day.
- At the poles, during solstice, the earth is either exposed to sunlight over the entire (24-hours) day or is completely hidden from the Sun throughout the entire day.
- This is why the **poles** get
  - no incoming radiation during their respective winter
  - more than the maximum radiation at the equator during their respective summer despite of smaller incidence angles of the Sun.

3. Solar radiation



www.polartrec.com

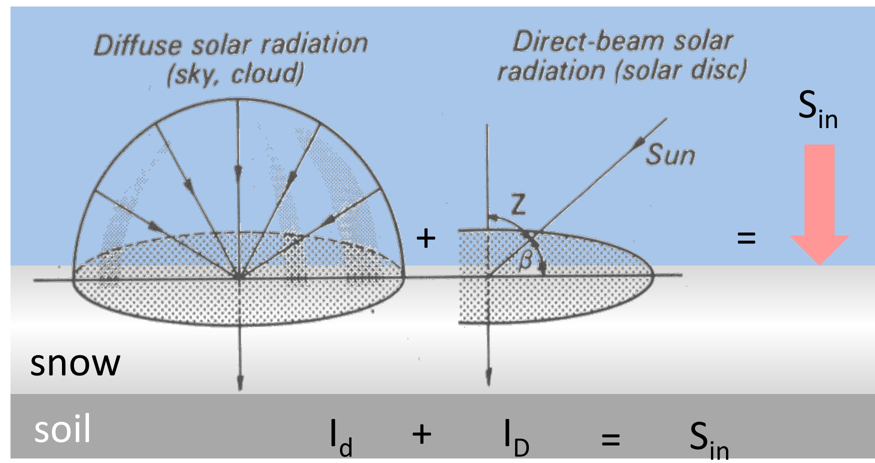
Lambert's cosine law

- Lambert's cosine law states that the energy flux density on a plane surface is directly proportional to the cosine of the incidence angle.
- The incidence angle h is the angle between the sun direction and the surface's normal.
- Since the incidence angle of the solar beam striking the horizontal ground is equal to sun the zenith angle Z, then:

$$I_D = I_{DN} \bullet \cos(Z)$$

And the global radiation on a horizontal surface can be expressed as:

$$S_{in} = I_{DN} \bullet \cos(Z) + I_d$$



S_{in}	**global radiation** flux (irradiance) on a horizontal surface [W/m²]
I_D	**direct** (horizontal) **radiation** flux on a horizontal plane [W/m²]
I_{DN}	**direct normal radiation** flux (beam irradiance) on a surface perpendicular to the direct beam [W/m²]
I_d	**diffuse radiation** flux (irradiance) [W/m²]
Z	Sun's zenith angle

Global radiation in complex topography

- In complex topography diffuse radiation originates from two sources: the sky and the surrounding topography and consists of three components:

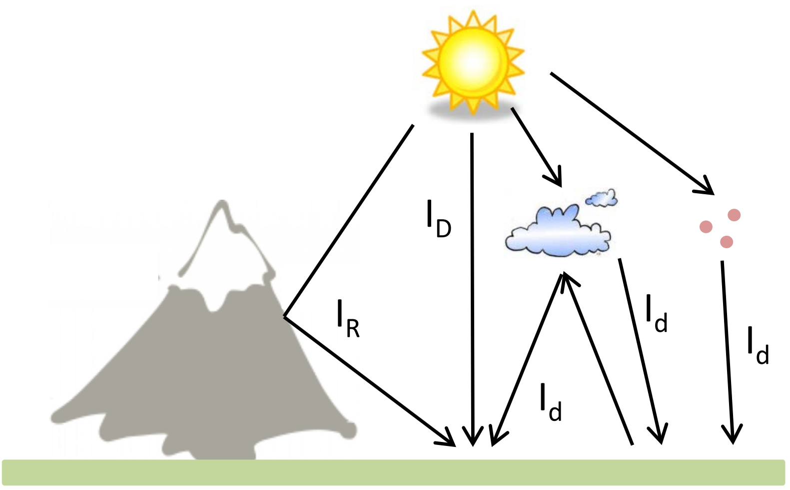
  1. radiation that is initially scattered out of the beam on the way through the atmosphere
  2. backscattered radiation e.g. global radiation that is reflected by the surface and subsequently redirected downward by scattering and reflection in the atmosphere.
  3. radiation reflected from adjacent slopes

- So the equation

$$\boxed{S_{in} = I_D + I_d}$$

can be extended by a term for reflected radiation (I_R):

$$\boxed{S_{in} = I_D + I_d + I_R}$$



I_R	radiation reflected off surrounding surfaces before arriving at the surface [W/m²]
I_D	direct (horizontal) radiation [W/m²]
I_d	diffuse sky radiation [W/m²]

Energy Balance

Forms of thermal energy transfer in the climate system

Energy moves in the climate system from one form to another. Thermal energy or heat can move from one place to another in three different forms:

Radiation

The transfer of energy through ***electromagnetic waves***. This form of energy transfer does not require the presence of matter to occur. In this form energy can travel through empty space from the Sun to the Earth. Radiation also occurs within the climate system between the earth's surface and the atmosphere, and within the atmosphere and ocean.

Advection/convection

The spreading of heat in fluids and gases through the ***flow of matter*** from one place to another. Advection occurs in the horizontal plane and convection in the vertical.

Conduction

The spreading of heat through ***molecular vibrations***. This form of heat transfer requires the presence of matter and can occur in solids, liquids, and gases. In the climate system conduction occurs mainly over small distances. In the atmosphere and oceans it is taken over by advection and convection.

Contents

4. Radiation measurement

Ground measurements of solar radiation

- Radiation measuring instruments can be classified according to their use.
- The generic term for all radiation measuring instruments is the radiometer.
- **Overview of radiometers**:

Name		Used for measuring
Pyrheliometer		direct solar beam
Pyranometer		global solar radiation (direct and diffuse)
Albedometer		difference between incoming and reflected solar radiation (albedo)
Pyrgeometer		longwave radiation
(Net) Pyrradiometer		difference between incoming and outgoing total radiation

4. Radiation measurement

ISO and WMO Specification and Classification

of instruments for measuring hemispherical solar and direct solar radiation

- There are three accepted categorization standards for solar instrument quality.
- These standards are defined differently by **ISO 9060:1990** and the World Meteorological Organization (**WMO**).

ISO 9060 Specification	WMO Classification	Application
Secondary Standard	high quality	reference instruments
First Class	good quality	operational networks
Second Class	moderate quality	low cost networks

4. Radiation measurement

Pyrheliometer (PHM)

- Measure direct beam solar radiation
- The general design of PHMs is a metal tube with a small opening at one end for the solar radiation.
- PHMs are ***oriented toward the Sun*** so the receptor surfaces are perpendicular to the incident solar beam.
- PHMs have an aperture with an ***acceptance angle of 2.5°-5°*** to limit the view to the solar disk.
- A PHM has to be attached to a mounting that permits it to follow the Sun (expensive).
- PHMs are the most accurate of all radiation instruments.
- ***MS-56 First Class Pyrheliometer*** has a ***response time < 1s*** and an excellent temperature stability (-40 °C to 80 °C)
- PHMs are commonly used as calibration standards for working instruments
- Usually found only at research station or laboratories because of their ***need to track the Sun***

DR01 Pyrheliometer

www.hukseflux.com

MS-56 Pyrheliometer

www.eko-eu-com

4. Radiation measurement

Pyranometer

- Instrument for measuring global radiation (direct and diffuse shortwave radiation), onto a plane surface
- A horizontally mounted pyranometer detects radiation from all parts of the sky.
- An inverted pyranometer measures reflected solar radiation.
- The most popular instruments are based on ***thermopiles*** that measure the thermal difference between a black and a white surface.
- The thermopiles are protected by one ore two ***hemispherical glass domes***, which provide isolation from the longwave radiation.

Star Pyranometer

www.logotronic.at

Kipp&Zonen

4. Radiation measurement

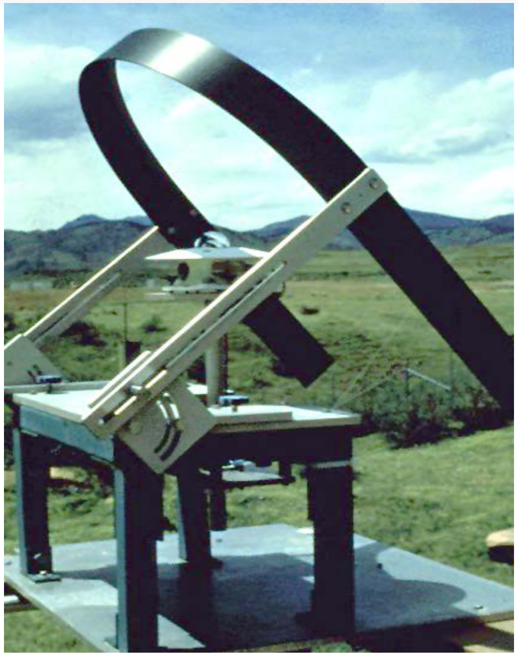
Pyranometer

- An upward facing pyranometer fitted with a ring device (occulting band) measures diffuse radiation, because the ***occulting band*** shades the sensor from the direct radiation.

- Many weather stations simply use two horizontal pyranometers - one fully exposed to the sky and the other shaded.

- For accurate measurements a pyranometer needs ***daily maintenance***.

- The dome must be cleaned every morning because ***dust, rain, snow, dew, glazed frost*** or ***hoar frost*** can be deposited on the top of the dome an can stay there.


www.nrel.gov


Eppley Laboratory, Inc

4. Radiation measurement

Ref No.	ISO-9060 Pyranometer Specifications	Secondary Standard	First Class	Second Class
1	**Response time:** time to reach 95% response	< 15 sec.	< 30 sec.	< 60 sec.
2	**Zero-offset:** Offset-A: response to 200 W/m² net thermal radiation, ventilated Offset-B: response to 5 K/h change in ambient temperature	+ 7 W/m² ± 2 W/m²	+ 7 W/m² ± 2 W/m²	+ 7 W/m² ± 2 W/m²
3a	**Non-stability:** % change in responsivity per year	± 0.8%	± 1.5%	± 3%
3b	**Non-Linearity:** % deviation from responsivity at 500 W/m² due to change in irradiance from 100 – 1000 W/m²	± 0.5%	± 1%	± 3%
3c	**Directional response (for beam irradiance):** the range of errors caused by assuming that the normal incidence responsivity is valid for all directions when measuring from any direction, a beam radiation whose normal incidence irradiance is 1000 W/m²	± 10 W/m²	± 20 W/m²	± 20 W/m²
3d	**Spectral Selectivity:** % deviation of the product of spectral absorbance and transmittance from the corresponding mean, from 0.35 – 1.5 µm	± 3%	± 5%	± 10%
3e	**Temperature response:** % deviation due to change in ambient within an interval of 50K, *(e.g. -10 to +40º C typical)*	2%	4%	8%
3f	**Tilt response:** % deviation in responsivity relative to 0º tilt, due to change in tilt from 0º to 90º tilt at 1000 W/m² beam irradiance	± 0.5%	± 2%	± 5%

Hukseflux, USA

4. Radiation measurement

Albedometer

- An albedometer is a combination of two pyranometers, one facing upward and one facing downward.
- The upward facing pyranometer measures global radiation (diffuse and direct solar radiation), while the downward facing pyranometer measures reflected solar radiation.

www.novalynx.com

- The Albedo is the fraction of incident solar radiation reflected by a surface.
- It can be calculated from the output data of an albedometer :

albedo = reflected radiation / global radiation

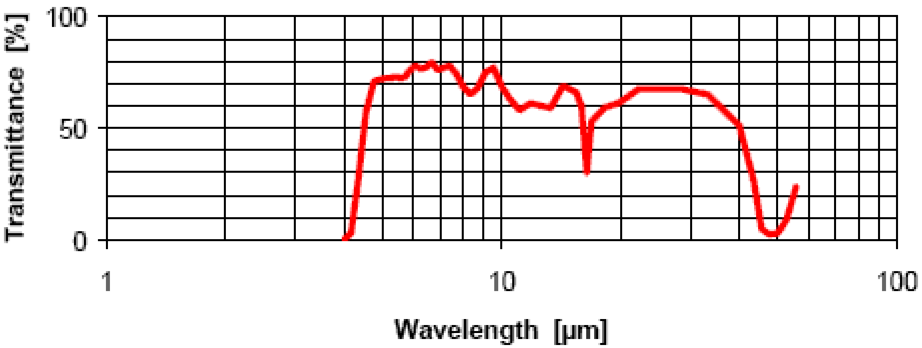
- Short-wave net radiation can also be calculated.

www.lambrecht.net

4. Radiation measurement

Pyrgeometer

- Instrument for measuring longwave radiation.
- A pyrgeometer resembles a pyranometer, but – instead of a glass dome - it has a ***polyethylene*** or ***silicon dome*** with good transmittance for the spectral domain of thermal IR in the range of 3.0 to 50 µm.
- Polyethylene or silicon ***domes filter out unwanted*** shortwave ***solar radiation***
- The reliability of a pyrgeometer is closely related to the transmittance of the dome, which must be ***cleaned daily***.
- Polyethylene domes must be replaced frequently.



Kipp & Zonen

4. Radiation measurement

Pyrradiometer (Net Radiometer)

- Net pyrradiometers measure the net total radiation flux (shortwave and longwave) downward and upward through a horizontal surface.
- ***Two*** black, radiation absorbing plates act as ***sensors***, one facing upward and one facing downward.
- Two separately working receivers.
- Each blackened disk has an internal thermopile, an the temperature difference between the two sensors is proportional to net radiation.
- Most pyrradiometers use ***polyethylene domes,*** because it is transparent to both shortwave and longwave radiation (0.3 to 100 µm)

www.novalynx.com

4. Radiation measurement

Measuring Net Radiation

- Net radiation is an important component of the surface energy budget.
- But net radiation remains one of the most difficult atmospheric parameters to measure.
- Each of the terms on the right side of the equation must be measured:

Two possible concepts:

1. A pair of pyranometers and a pair of pyrgeometers.

   One instrument from each pair facing upward and one downward.

2. Pyrradiometer (net radiometer)

Net radiation Q_{NR} is the sum of net shortwave S and longwave radiation L:

$$\boxed{\begin{aligned} Q_{NR} &= S_{net} + L_{net} \\ &= (S_{in} - S_{out}) + (L_{in} - L_{out}) \end{aligned}}$$

S_{net}	net incoming shortwave radiation
L_{net}	net incoming longwave radiation
S_{in}	incoming shortwave radiation
S_{out}	reflected shortwave radiation
L_{in}	incoming longwave radiation
L_{out}	outgoing longwave radiation

4. Radiation measurement

Sunshine recorder

- Instrument for the measurement of sunshine duration.
- WMO definition of sunshine: Duration of the period for which the *direct* solar irradiance exceeds **120 W/m²**.
- This value is equivalent to the level of solar irradiance shortly after sunrise or shortly before sunset in cloud-free conditions.

Campbell-Stokes sunshine recorder

Correlation between sunshine duration and global radiation (***Angström-Formula***):

$$G / G_0 = a + b \cdot (SD / SD_0)$$

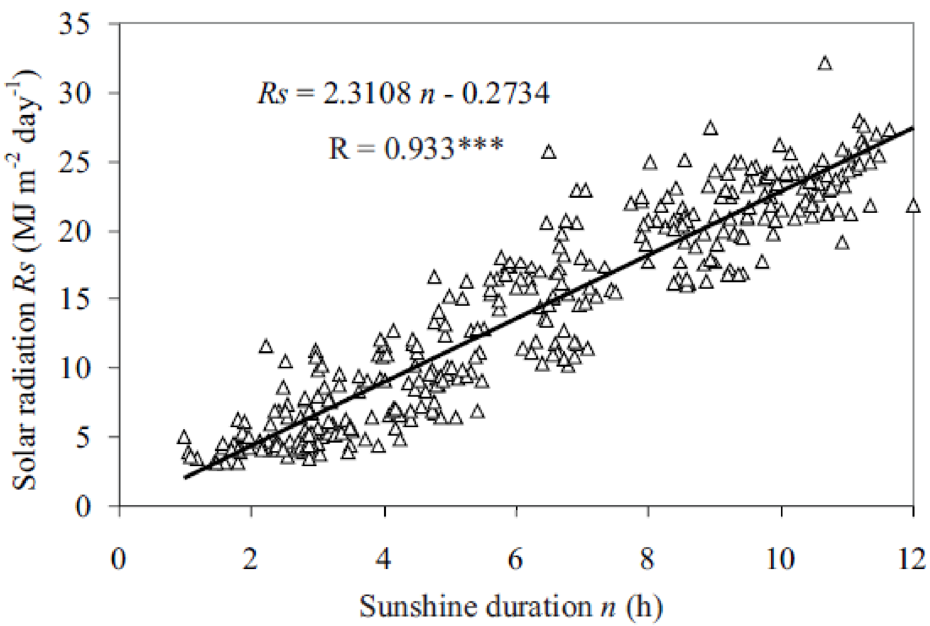
G	global solar radiation (S_{in})
G_0	extra-terrestrial global radiation
SD	sunshine duration
SD_0	extra-terrestrial potential SD
a, b	constants which have to be determined monthly

All parameters of the Angström-Formula are ***monthly averages*** of daily values

4. Radiation measurement

Correlation between Sunshine Duration and Global Radiation

- Knowledge of global solar radiation is essential in the prediction, study and design of the economic viability of systems which use solar energy.
- In many countries sunshine duration is measured at more locations than the global radiation.
- The Angström-Formula can then be used to estimate global solar radiation for places where only sunshine records are available.



Correlation between global radiation (*G*) and sunshine duration (*SD*) in Constanta, Romania, 1971-2000, mean monthly values.

For mean monthly values we often find good correlation between G and SD.

4. Radiation measurement

Sunshine duration

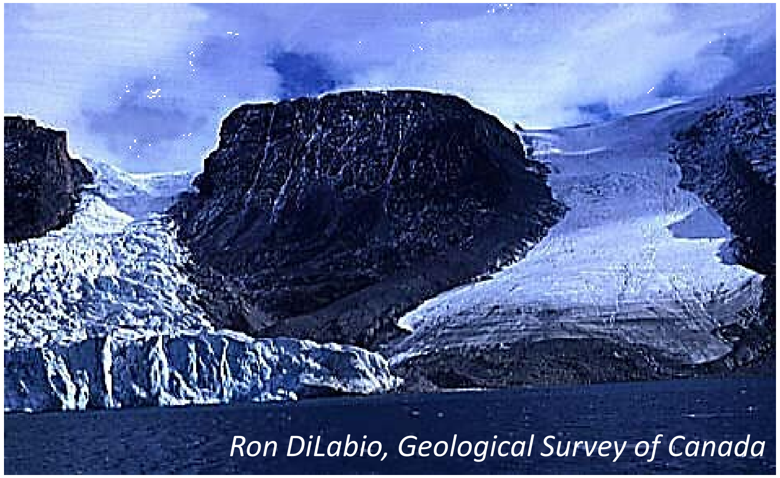
Based on many measurements made at various locations on the Earth and published by more than one author, Allen *et al.* [1] recommended the values of $a = 0.25$ and $b = 0.50$ in estimating *global radiation G*, when there is available data on sunshine duration and direct measurements on *G* are missing

Contents

5. Characteristics of snow and ice

Introduction

- Ice and snow melt at 0°C, but this does not necessarily mean that melting will occur with an air temperature > 0°C or that there is no melting at an air temperature < 0 °C.
- Snow and glacier melt are determined by the ***energy balance*** at the snow/glacier-atmosphere interface, for which air temperature is only an index.
- The energy balance is controlled by
  - the ***meteorological conditions*** above the snow/glacier surface.
  - the ***physical properties*** of snow/glacier surface.



Ron DiLabio, Geological Survey of Canada

There is a large spatial heterogeneity of the energy balance components, typically encountered in steeply sided terrain.

5. Characteristics of snow and ice

Special Characteristics of Snow and Ice

relevant for melt

Fixed surface **temperature** (**O °C**) during melting

Vapour pressure of a melting surface: **6.11 hPa**

High and largely variable **albedo**

High **thermal emissivity**

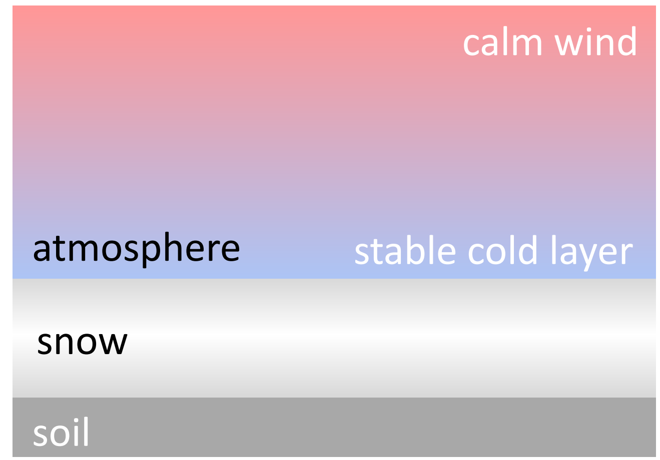
Penetration of shortwave radiation

Generally low **surface roughness**

5. Characteristics of snow and ice

Surface temperature of snow and ice

- The surface temperature over snow and ice cannot exceed **0 °C:**
  - ***strong temperature gradients*** can develop in the air immediately above the surface.
  - temperature gradients may reach more than **5 °C/m** within the first 2 m above the surface.
  - during melt season the **air** is generally **stably stratified,** i.e. strong **temperature inversion** close to the surface.
  - without wind, the stable layer does not break up and ***turbulent fluxes*** become ***minimal***.

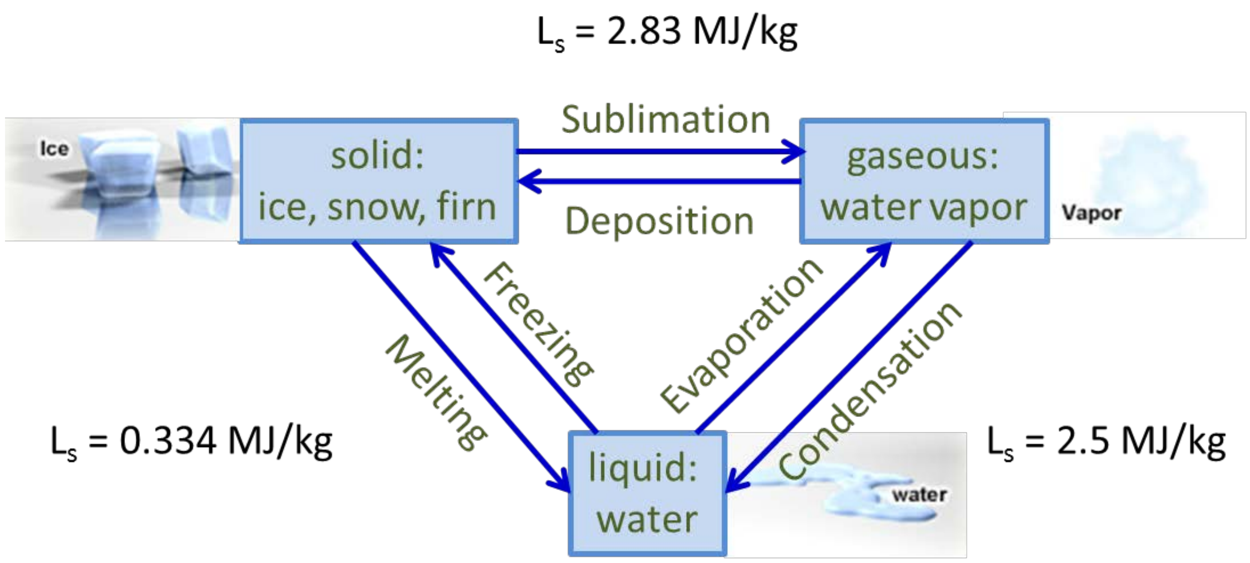


- During winter and cold nights the snow or ice ***surface*** can be ***warmer than air:***
  - energy losses from the surface to the air, due to evaporation
  - significant reduction of the energy available for melt

5. Characteristics of snow and ice

Vapor pressure over snow and ice

- The vapour pressure over snow and ice cannot exceed **6.11 hPa.**
- This rel. low value favours vapour pressure gradients towards the surface ➡ **condensation**
- Since the latent heat of evaporation is **7.5** times larger than the latent heat of fusion, condensation can be an important energy source for melt.



5. Characteristics of snow and ice

Vapor pressure over snow and ice

CONDENSATION over a melting snow or ice surface

- **increase** of releasing **latent heat**
- **increase** of **moisture** of the snow/ice surface
- often **decrease** of the **albedo**
- **increase** of the **net shortwave radiation**

The higher the air humidity, the **higher** the **net longwave radiation**

increase of available **melt energy**

5. Characteristics of snow and ice

Reflectance (Albedo)

- Shortwave radiation is normally the main energy input for a snow cover.
- So albedo is one of the most important parameters driving snow energy balance.
- Albedo of Snow is generally higher than albedo of ice.

Albedo of fresh snow varies roughly between **0.7 and 0.9.**

Albedo of ice varies roughly between **0.3 and 0.5.**

Reflectance depends on:

Characteristics of snow:
- snow grain size
- snow crystal orientation
- snow crystal shape
- snow water content
- impurities e.g. black carbon

Albedo is highest for fresh snow, and decreases rapidly after a few days.

Characteristic of radiation:
- Angle of incidence - is highest at low angles of incidence
- higher for diffuse than for direct radiation
- radiation wavelength
- Cloudiness

5. Characteristics of snow and ice

Spatial variability of albedo of Aletsch Glacier, Switzerland.

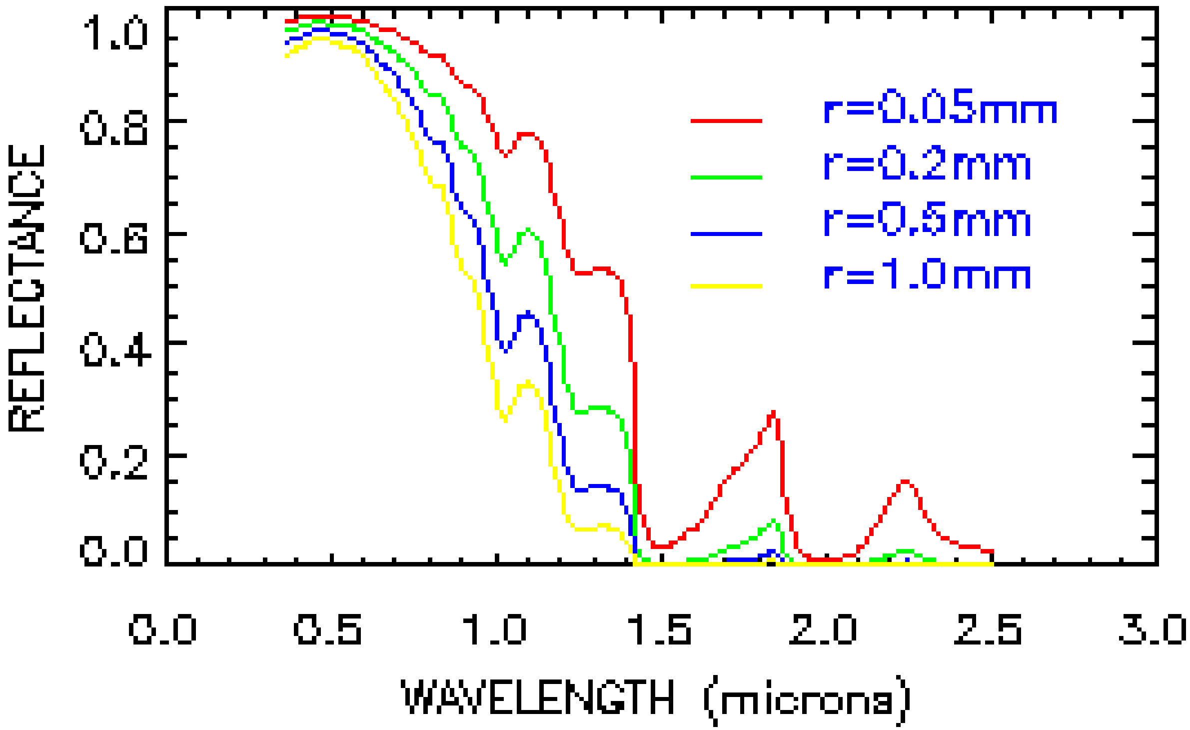
Source: GoogleEarth

5. Characteristics of snow and ice

Reflectance of Snow

- Snow reflectance is strongly dependent on wavelength:
  - ***highest*** in the ***visible spectrum*** (shortwave)
  - ***very low*** in the ***longwave region***

- Snow reflectance decreases as the snow grain size (r) increases.



5. Characteristics of snow and ice

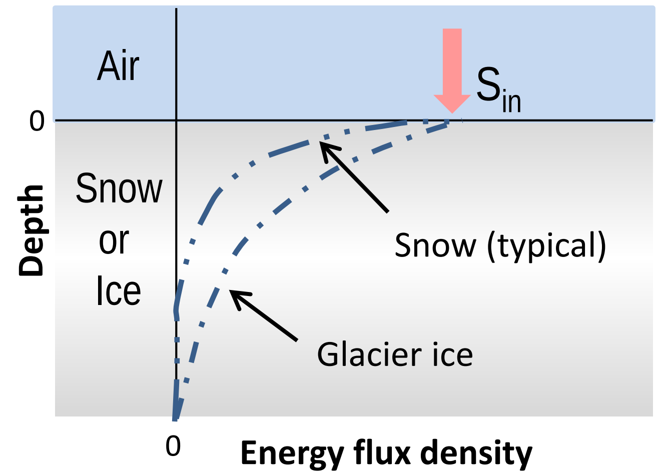
Emissivity

- In the infrared part of the spectrum snow and ice behave as almost perfect black-bodies:

Emissivity of:	
Snow:	0.98 -0.99
Ice:	~0.97
Soil:	0.95-0.97

Transmission

- Snow and ice allow some transmission of shortwave radiation.
- Only about ***1-2%*** of global radiation penetrates into a snow cover.
- Shortwave radiation penetrates ice and snow to a depth of about **10 m** and **1 m**.
- Transmission of radiation ***declines exponentially***.
- Most of the energy is absorbed in the first few cm below a snow surface.
- Is the height of ***a snow cover > 20 cm transmission*** is ***negligible***.



After Oke, 1987

5. Characteristics of snow and ice

Transmission

Transmission through snow or ice can be approximated by

Beer's Law for radiation penetration through a homogenous media

$$S_{in[z]} = S_{in[0]} \cdot e^{-vz}$$

$$z = -\ln\left(\frac{S_{in[z]}}{S_{in[0]}}\right) \Big/ v$$

$S_{in[z]}$	shortwave radiation at depth z [W/m^2]
$S_{in[0]}$	shortwave radiation at the surface [W/m^2]
v	extinction coefficient [m^{-1}]
z	depth [m]
$\frac{S_{in[z]}}{S_{in[0]}}$	fraction of shortwave radiation at depth z to shortwave radiation an surface [-]

Extinction coefficients [m^{-1}]

Low-density snow: 40
High-density snow : 10
Ice: 1

Contents

6. Parameterization of radiation

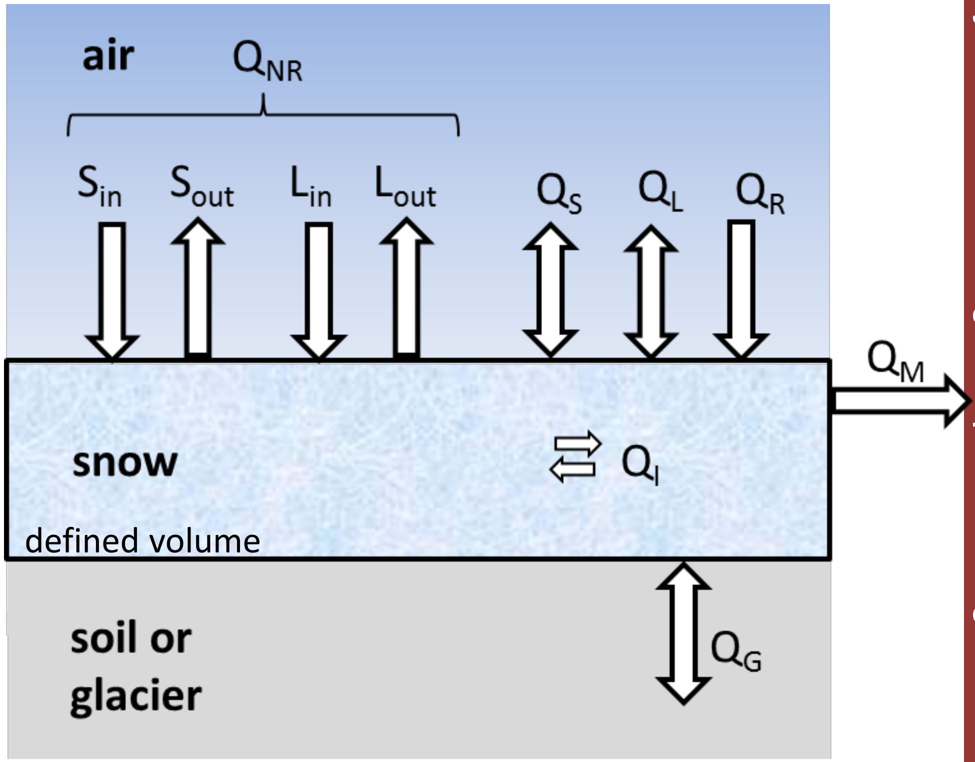
Energy Balance Equation

$$\frac{dU_I}{dt} = \sum Q_{in} - \sum Q_{out}$$

$$Q_I = Q_{NR} + Q_S + Q_L + Q_R + Q_G + Q_M$$

Take care with positive/negative signs!
Which fluxes are positive / negative?

U_I	snowpack internal sensible and latent heat storage
Q_{NR}	net radiant energy exchange
Q_S	sensible heat exchange with the atmosphere
Q_L	latent heat exchange of vaporization and sublimation with the atmosphere
Q_R	heat provided by rain
Q_G	heat from conduction in the ground
Q_I	change in snowpack internal sensible and latent heat storage
Q_M	Loss of latent heat of fusion due to meltwater leaving the snowpack



Schematic of the energy balance for a snowpack

Net radiation Q_{NR}

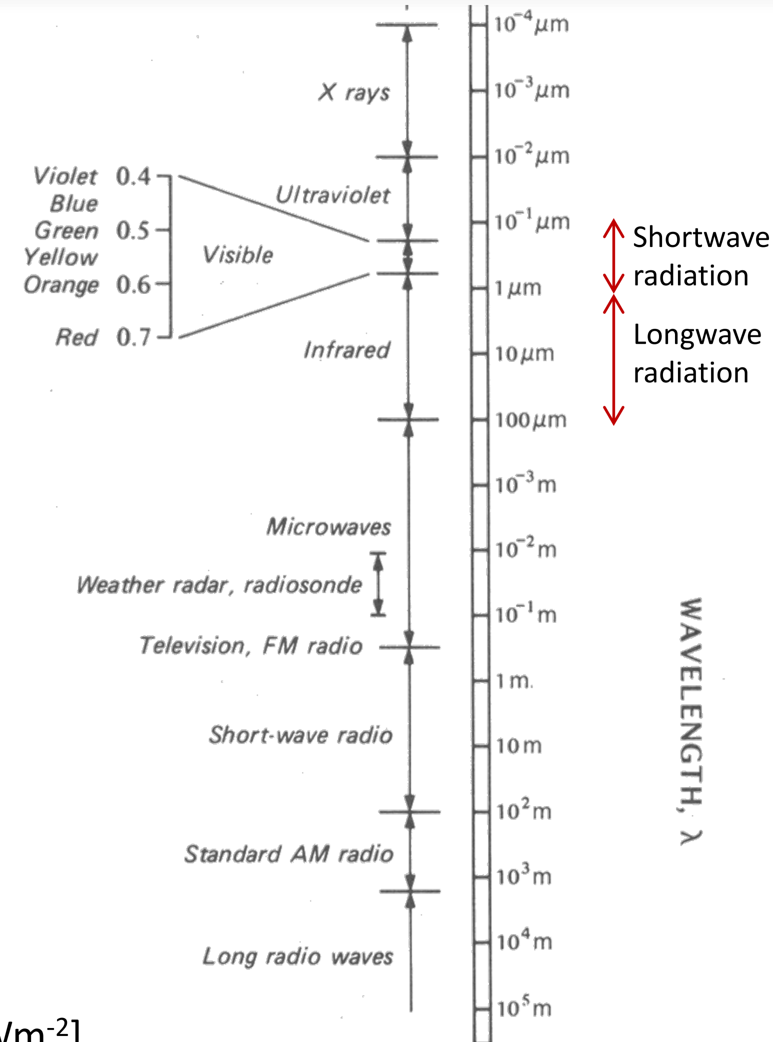
Net radiation Q_{NR} is the sum of net shortwave S and longwave radiation L:

$$\begin{aligned} Q_{NR} &= S_{net} + L_{net} \\ &= (S_{in} - S_{out}) + (L_{in} - L_{out}) \end{aligned}$$

S_{net}	net incoming shortwave radiation
L_{net}	net incoming longwave radiation
S_{in}	incoming shortwave radiation
S_{out}	reflected shortwave radiation
L_{in}	incoming longwave radiation
L_{out}	outgoing longwave radiation

Shortwave radiation: Wavelenth of 0.15 - 2 µm
Longwave radiation: Wavelenth of 2 - 100 µm

- ***Q_{NR}*** can be ***positive or negative***.
- Q_{NR} varies typically between about -100 and 300 [Wm^{-2}].
- During the night, when S_{net} is zero, Q_{NR} is entirely determined by L_{net}.



The electromagnetic spectrum (Oke, 1987)

Shortwave radiation

- ***Net shortwave radiation*** generally represents the ***major energy source*** for snow and glacier melt.

$$\boxed{S_{net} = S_{in} - S_{out}}$$

$$\boxed{S_{out} = \alpha \cdot S_{in}}$$

S_{net}	net incoming shortwave radiation
S_{in}	incoming shortwave radiation
S_{out}	reflected shortwave radiation
α	albedo

- Upon entering the atmosphere, solar radiation is portioned into **direct** and **diffuse** components.
- The ***proportion of diffuse*** radiation is:
  - about 10-20% under ***cloud-free condition***
  - 100% under ***complete cloud cover***

- The fraction of diffuse radiation is higher at lower solar altitudes and in the morning or in the evening

- Measurement of net radiation on glaciers are seldom available ➡ it is necessary to parameterize the individual components.

Direct solar radiation

- In mountainous regions, direct shortwave radiation ***varies considerably in space and time*** as a result of:
  - ***slope*** (reflection and emission of the surrounding slopes)
  - ***aspect*** (exposition)
  - ***effective horizon*** (obstruction of the sky)

Potential clear-sky direct solar radiation

$$I_c = I_0 \boxed{\left(\frac{R_m}{R}\right)^2} \cdot \psi_a^{\frac{p}{p_0 \cdot \cos Z}} \cdot \cos\theta$$

Iqbal, 1983

Eccentricity correction factor of the Earth's orbit

P/P_0 accounts for **effect of altitude**

I_c	potential clear-sky direct radiation on a inclined surface
I_0	solar constant (~1366 Wm^{-2})
R_m	mean Sun-Earth distance [m]
R	current Sun-Earth distance [m]
ψ_a	atmospheric clear-sky transmissivity
P	atmospheric pressure [hPa]
P_0	atm. pressure at sea level [hPa]
Z	local zenith angle
θ	angle of incidence between the slope-normal and the solar beam

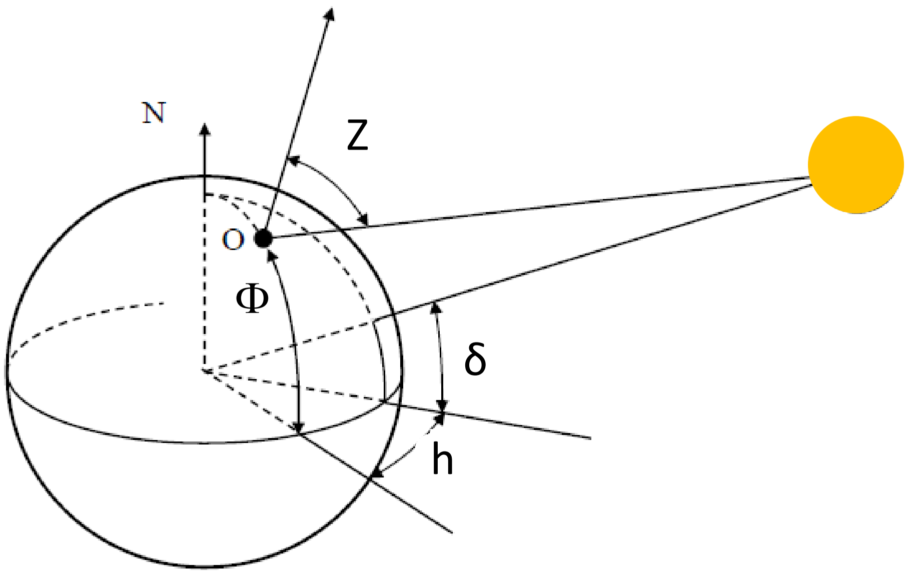
Direct solar radiation

Definitions:

Φ : **geographic latitude**:
north latitudes are positive, south latitudes are negative

Z : **zenith angle**:
angle between the vertical above the observer (the normal) and the sun.

δ : **solar declination**:
angle between the sun beam and the plane of the Earth's equator.



$\delta > 0$: when the sun's rays are north of the equator, i.e. for the summer period in the northern hemisphere, March 22 to September 22 approximately.

$\delta < 0$: when the sun's rays are south of the equator.

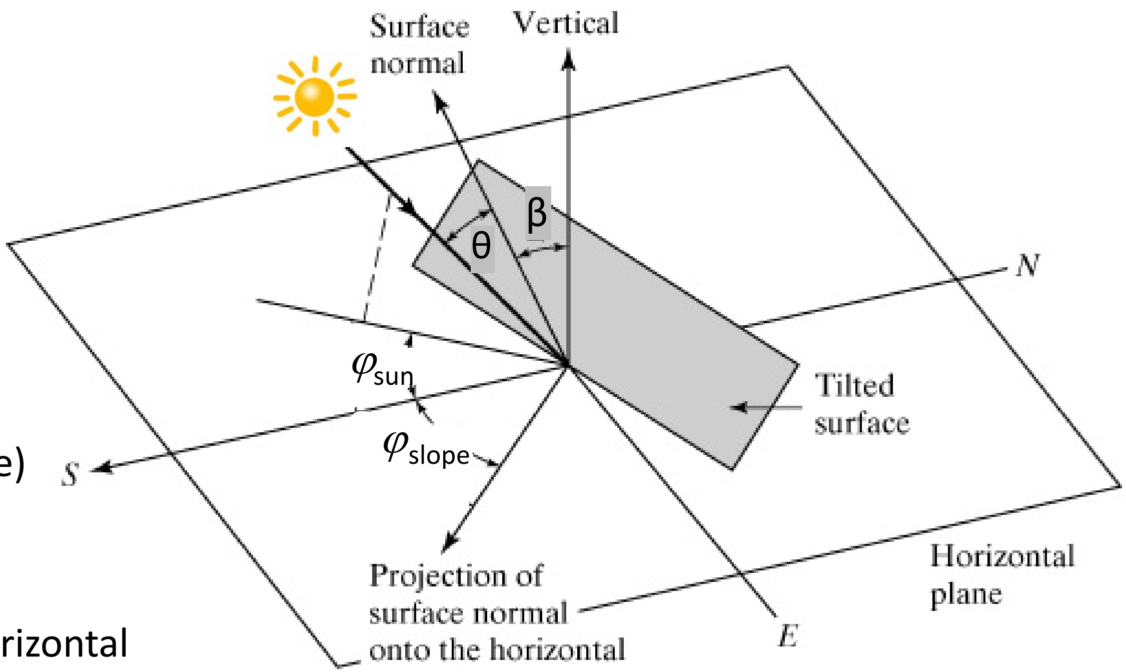
h : **hour angle**:
$h < 0$: before solar noon
$h > 0$: after solar noon

Direct solar radiation

Definitions:

θ : **incidence angle**:
angle between the solar beam and the slope normal.
For a horizontal surface
θ = Z (zenith angle)

β : **surface tilt angle**:
angle between the surface (slope) normal and the vertical.



φ_{slope}: **surface azimuth angle**:
angle between south and the horizontal projection of the surface normal.
negative: for a slope that faces east of south
positive: for a slope that faces west of south

φ_{sun}: **solar azimuth angle**:
angle in the horizontal plane measured from south to the horizontal projection of the sun beam.
negative: east of south
positive: west of south

Direct solar radiation

- A widely used solution for the incidence angle is (Garnier and Ohmura, 1968):

$$\cos\theta = \cos\beta \cdot \cos Z + \sin\beta \cdot \sin Z \cdot \cos(\varphi_{sun} - \varphi_{slope})$$

θ	angle of incidence between the slope normal and the solar beam
β	slope angle (surface tilt angle)
Z	local zenith angle
φ_{sun}	solar azimuth angle
φ_{slope}	slope azimuth angle

$$\cos Z = \sin\Phi \cdot \sin\delta + \cos\Phi \cdot \cos\delta \cdot \cos h$$

Z	local zenith angle
Φ	local latitude
δ	current declination of the sun
h	hour angle, in the local solar time

Direct solar radiation

$$\delta = -23.44 \cdot \left[360° \cdot \frac{(N+10)}{365} \right]$$

δ	current declination of the sun
N	julian day number (1= january 1)

The number 10, in (N+10), is the approximate number of days after the December solstice to January 1.

$$h = 15 \cdot (t - 12)$$

h	hour angle, in the local solar time
t	local solar time. It is based on a 24-hour clock, with 12:00 as the time that the sun is exactly due south.

- The hour angle is equal to zero at true solar noon, increasing by 15° per hour.
- **h < 0**: before solar noon
- **h > 0**: after solar noon

Direct solar radiation

- As a quick reminder:

Potential clear-sky direct solar radiation

$$I_c = I_0\left(\frac{R_m}{R}\right)^2 \cdot \psi_a^{\frac{p}{p_0 \cdot \cos Z}} \cdot \cos\theta$$

Iqbal, 1983

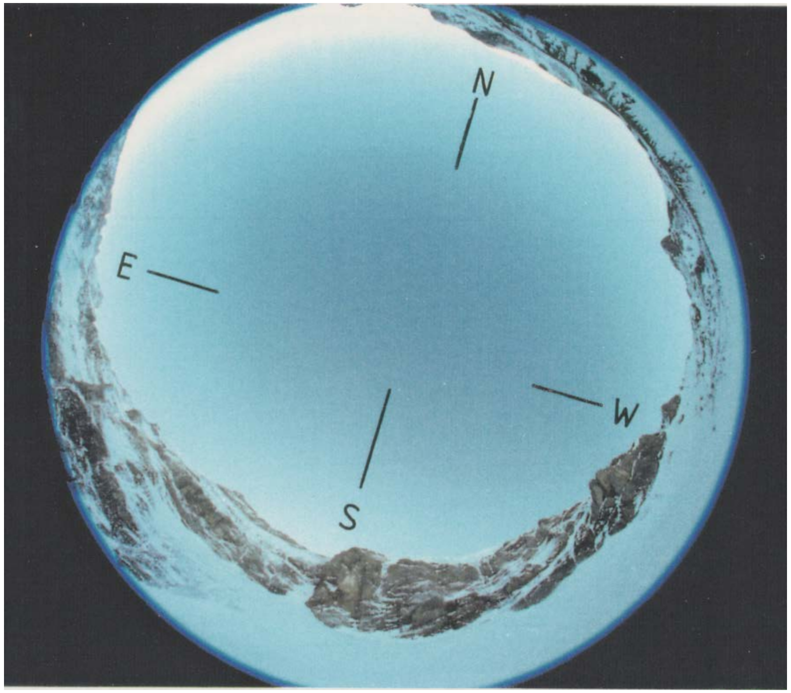
- The ***atmospheric transmissivity*** ψ_a varies considerably from place to place according to the weather and air mass conditions.
- Transmissivity tends to:
  - be ***higher in winter*** and ***lower in summer***
  - ***increase with latitude***

due to the lower atmospheric water vapour and dust content both in winter and in high latitudes.

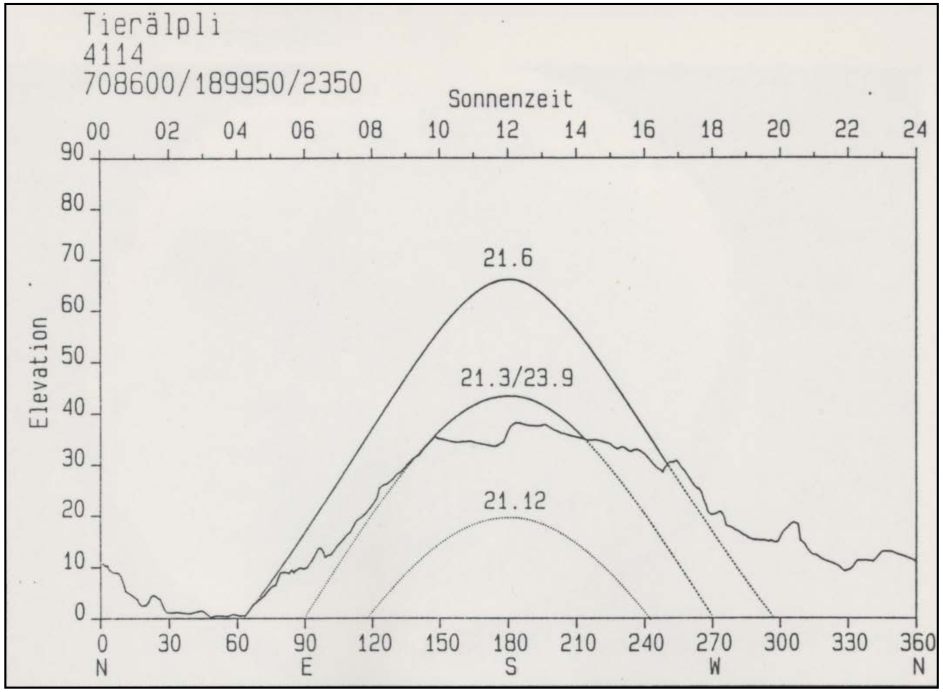
- Clear-sky transmissivities typically vary ***between 0.6 and 0.9***.

Shading at a particular point

Fisheye photography



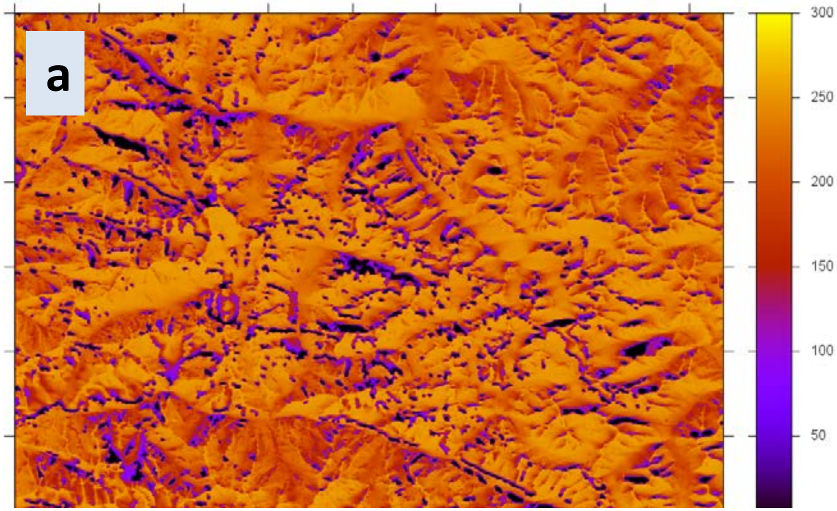
Elevation, sun path and shading of the sun



6. Parameterization

Shortwave radiation

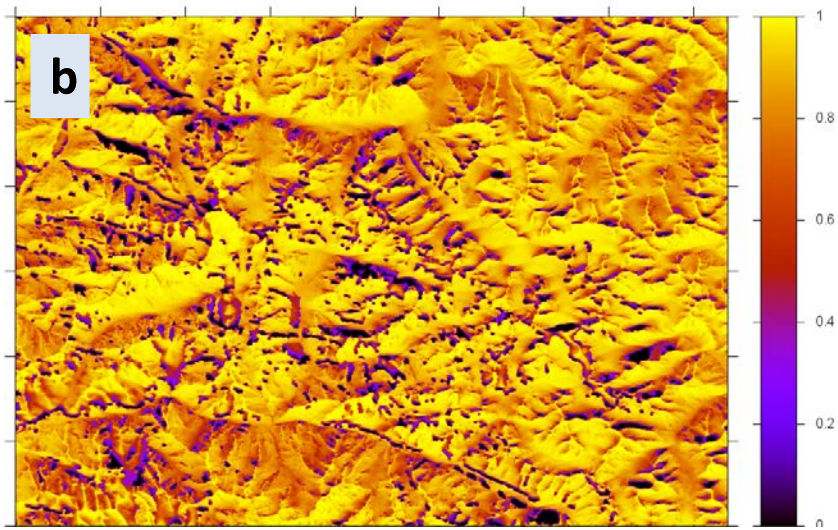
Shading for a whole region (DTM)

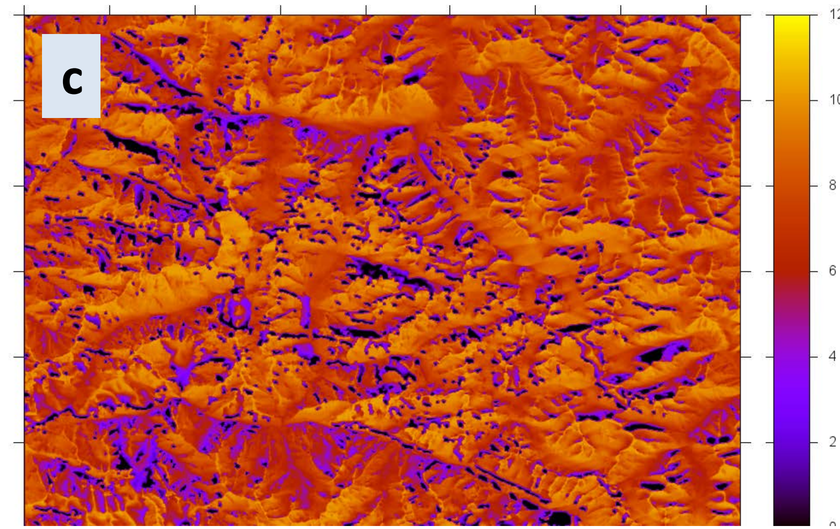


Potential (extra-terrestrial) solar radiation on a horizontal plane

Central Himachal Pradesh, Nov. 11

a. Potential solar radiation [W m^{-2}]
b. Fraction of potential solar radiation [%]
c. Potential sun shine duration [h]



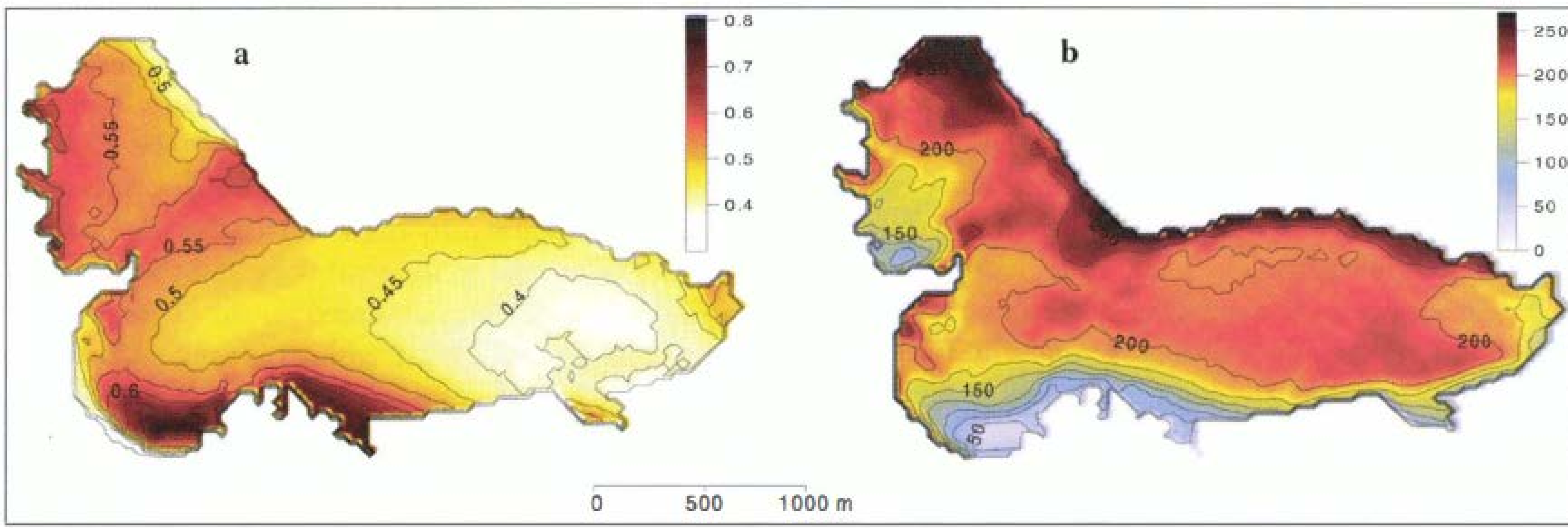


Spatial variation of shading

Effects of topography on shading and potential direct solar radiation for Storglaciären, Sweden

a) Ratio of number of shaded hours to total number of hours

b) Potential clear-sky direct solar radiation [W m^{-2}]



(DEM , 30 m resolution, slope angle considered)

Hock, 1999

6. Parameterization

Different temperature-index methods for melt modelling

Temperature-index methods

Classical degree-day factor

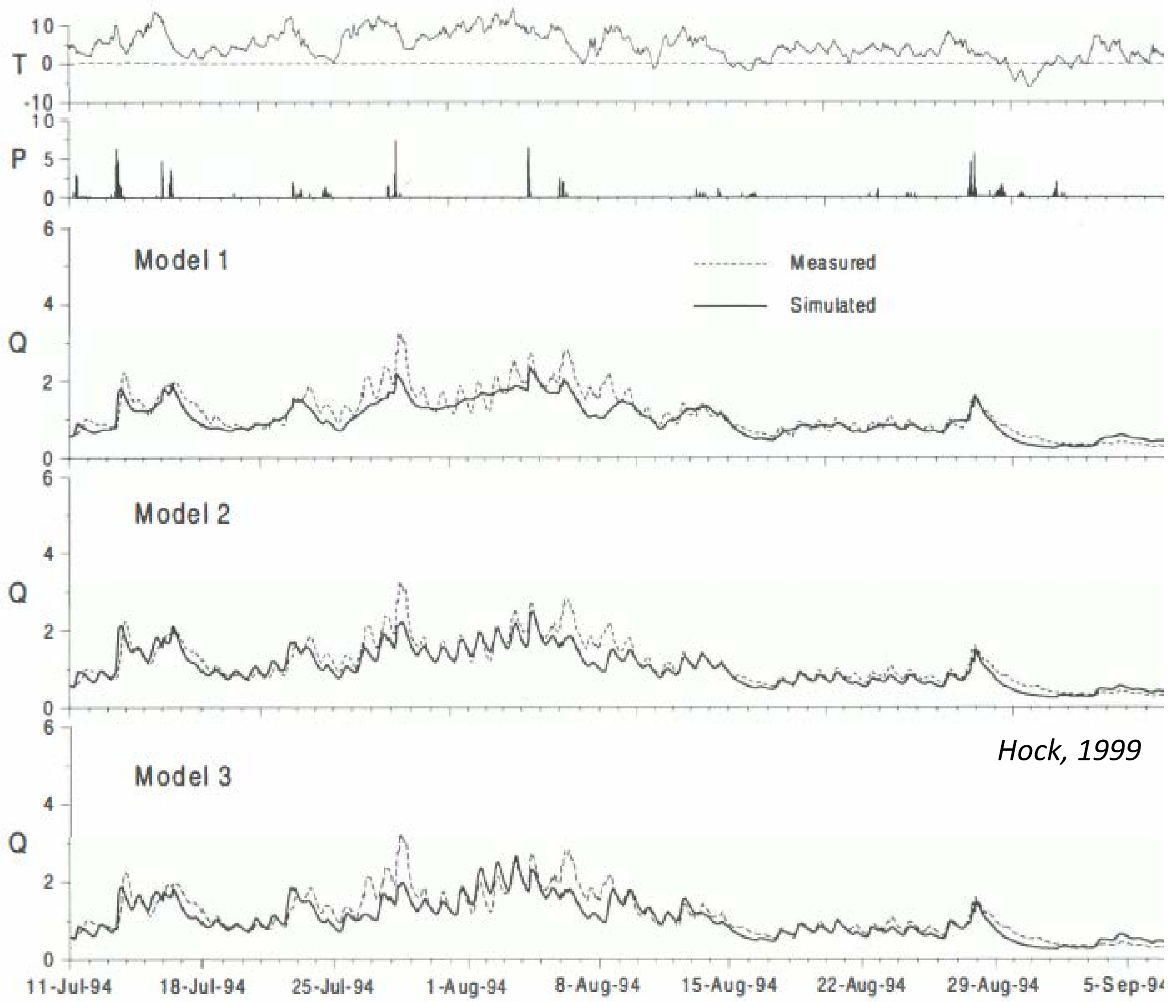
- constant in space and time
- $DDF_{ice} \neq DDF_{snow}$

Including potential clear-sky direct solar radiation

- spatial and daily variability
- radiation-index in terms of pot. direct solar radiation
- topographic effects considered (slope, aspect, effective horizon)

Including potential clear-sky direct solar radiation and measured global radiation

- global radiation data, measured in the area needed
- cloudy and overcast conditions considered



Hock, 1999

Diffuse solar radiation

- The amount of diffuse radiation depends largely on atmospheric conditions.
- In complex topography ***diffuse radiation originates*** from:
  - sky
  - surrounding topography

- It consists of ***three components***:
  - **initially scattered radiation**: sky radiation:
  - **backscattered radiation**: global radiation that is reflected by the surface and subsequently redirected downward in the atmosphere mostly by clouds
  - **reflected radiation**: global radiation that is reflected from adjacent slopes

- Surrounding topography affects the amount of diffuse radiation in two opposing ways:
  - **positive**: sky radiation is reduced, as part the sky is obscured.
  - **negative**: diffuse radiation is enhanced by reflection from adjacent slopes.

6. Parameterization

Shortwave radiation

Diffuse solar sky radiation

- A surface, whether in shadow or not, can receive diffuse sky radiation only if the portion of the sky of the overlying atmosphere is unobstructed.
- In many cases, diffuse sky radiation is parameterized using a so called ***sky view factor***.

$$\boxed{V_f = \cos^2(H)}$$

- A widely used simplification is:

$$\boxed{V_f = \cos^2\left(\frac{\beta}{2}\right)}$$

H	average horizon angle
β	slope angle (surface tilt angle)
V_f	sky view factor

- The ***sky view factor*** V_f is related to the fraction of the hemisphere unobstructed by surrounding slopes.
- $\mathbf{V_f = 1}$: the sky is ***completely unobstructed***
- $\mathbf{V_f = 0}$: the sky is ***totally obstructed*** and no diffuse radiation reaches the surface
- Fisheye-photos are one experimental possibility for determining the sky view factor.

Diffuse solar radiation

- Diffuse radiation on a ***surface*** including the ***effects of topography*** can be approximated by:

$$D = D_0 \cdot V_f + \alpha_m \cdot G(1 - V_f)$$

sky radiation **terrain** radiation

D	Diffuse radiation
D_0	Diffuse sky radiation of an unobstructed sky
G	global radiation
V_f	sky view factor
α_m	mean albedo of the surroundings

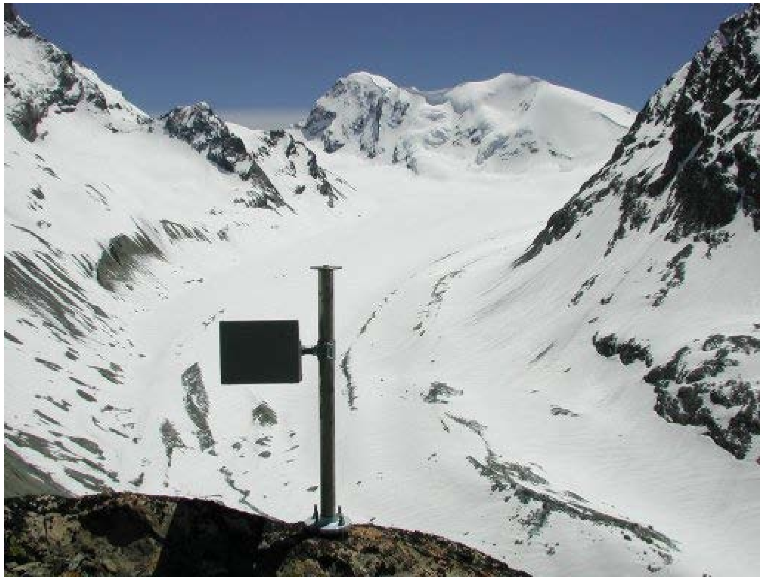
- D ~ 10-20% under ***clear sky condition***
- D = 100% under ***complete cloud cover***

6. Parameterization

Estimating albedo using terrestrial photographs

- Since shortwave radiation plays a major role in the energy budget of temperate glaciers, the albedo is one of the leading variables controlling the energy balance.
- Consequently, an accurate estimation of the surface albedo is essential.
- One possible method to estimated the temporal and spatial distribution of the albedo is using automatic cameras.

Haut Glacier d'Arolla, Switzerland



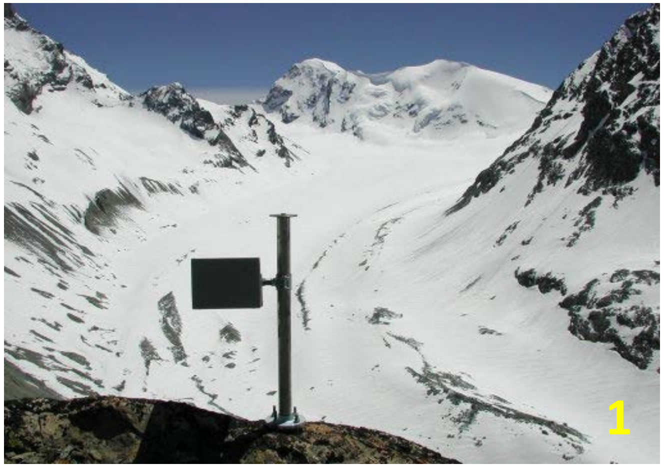
http://www.arolla.ethz.ch/georef/arollanimation.html

Petermann Glacier, Greenland

http://www.agu.org

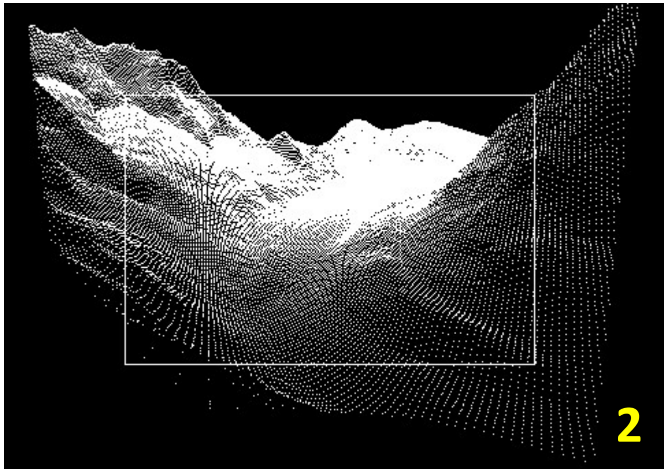
Estimating albedo using terrestrial photography

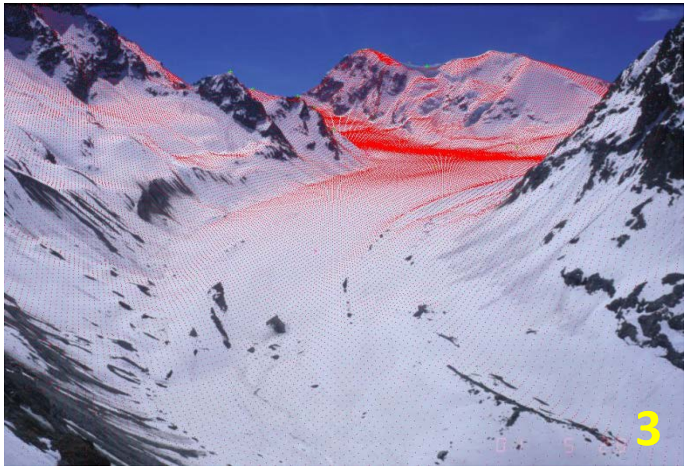
The method allows to estimate the temporal and spatial variations of surface albedo, as well as the fraction of snow covered area, on glacier and snow-covered mountainous areas.



Advantage: inexpensive and practical method
Shortcoming: requires at least one measurement of albedo to be made in the field at a location visible in the image to estimate the albedo in the remaining pixels.

1) Photograph of a digital camera
2) Digital elevation model (DEM)
3) Georeferenced photograph





Pictures: Corripio, J.G., ETH Zurich

Estimating snow cover evolution using terrestrial photography

- The following animation shows ***changes in the snow cover*** over the Haut Glacier d’Arolla.
- Series of georeferenced ***maps of reflectance values***
- The values are derived by georeferencing some of the ***daily photographs*** taken by an ***automatic camera***.

6. P… wave radiation

Estimating albedo

- A simple approach to modelling snowpack albedo is:

$$\boxed{\alpha = \alpha_0 + K \cdot e^{-nr}}$$

α	snowpack albedo
α_0	minimum snowpack albedo ~ 0.4
K	constant ~ 0.44
n	number of days since last major snowfall
r	recession coefficient

- This parameterization leads to good average results over longer time periods (weeks, months).
- The day to day variations are not always well represented.

Longwave radiation

- Unlike shortwave radiation exchange that is restricted to daytime, ***longwave radiation exchange*** occurs both ***day and night***.
- L_{net} can be ***positive or negative***.

$$\boxed{L_{net} = L_{in} - L_{out}}$$

$$\boxed{L_{net} = \varepsilon_a \cdot \sigma \cdot T_a^4 - \varepsilon_s \cdot \sigma \cdot T_s^4}$$

L_{net}	net incoming longwave radiation
L_{in}	incoming longwave radiation
L_{out}	outgoing longwave radiation
σ =	Stefan-Boltzmann constant = 5.67 x 10^{-8} [W m^{-2} K^{-4}]
T_a	temperature of the air [K]
T_s	temperature of the snow/ice surface [K]
ε_a	emissivity of the air [-]
ε_s	emissivity of the snow/ice surface [-]

- In ***most cases***, L_{out} exceeds L_{in} i.e., ***L_{net} is negative***.
- However, L_{net} becomes less negative and can even be positive under heavy cloud cover, forest cover, and fog.

Longwave radiation

Longwave incoming radiation (L_{in})

- L_{in} is emitted by atmosphere - mostly by ***water vapour*** *(~81%)*, ***CO₂*** *(~17%)* and ***O₃*** *(~2%).*
- The most important part of L_{in} is ***produced in the lowest 1-2 km*** of the atmosphere.
- ***Variations are largely*** due to variations in cloudiness and the temperature of water vapour.
- Higher air temperature, water vapour and more clouds -> higher L_{in}.
- L_{in} tends ***decrease with altitude***.
- L_{in} varies between about 250 and 350 [Wm^{-2}]

In mountainous areas:

- Surrounding topography can cause significant spatial variations in L_{in}.
- L_{in} is ***reduced by obstructed sky*** due to surrounding terrain.
- Conversely, the surface receives additional radiation from the surrounding terrain and the air between the terrain and the receiving surface.

Longwave radiation

Longwave incoming radiation (L_{in})

- In melt models, L_{in} is usually ***estimated from empirical relationships*** based on standard meteorological measurements.
- These relationships exploit the fact that L_{in} ***correlates*** well ***with air temperature and vapour pressure*** at screen level.
- Generally, the equations for these approaches take the form:

$$\boxed{L_{in} = \varepsilon_{ac} \cdot \sigma \cdot T_a^4 \cdot F(N)}$$

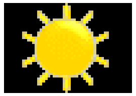
L_{in}	incoming longwave radiation
σ =	Stefan-Boltzmann constant = 5.67×10^{-8} [W m^{-2} K^{-4}]
T_a	temperature of the air [K]
ε_{ac}	clear sky emissivity [-]
$F(N)$	cloud factor (function of cloud amount N)

- The **cloud factor F(N)** describes the increase in radiation due to clouds as a functions of the amount of clouds.
- Cloud cover significantly increases longwave atmosphere radiation, as ***clouds*** are characterized by a ***high emissivity very close to 1***.

Longwave radiation

Longwave incoming radiation (L_{in})

- Many approximately equivalent forms of empirical equation for clear sky atmospheric emissivity exist.
- The equation given by ***Brutsaert (1975)*** is often used because of its ***simplicity*** and the fact that it does ***not require calibration to local condition***:



Clear-sky conditions:

$$\varepsilon_{ac} = 1.24 \cdot \left(\frac{e_a}{T_a} \right)^{1/7}$$

ε_{ac}	clear sky emissivity
e_a	vapour pressure [hPa]
T_a	temperature of the air [K]

- Cloudy-sky atmospheric emissivity (εa) is obtained from clear-sky atmospheric emissivity defining an empirically derived cloud factor that describes the increase in radiation as a function of the ***fraction of cloud covered sky (N)***.
- One method to calculate cloudy-sky longwave radiation is simply obtained multiplying clear-sky longwave radiation by a ***factor that increases with increasing N***.

Overcast conditions:

$$L_{in} = \varepsilon_{ac} \cdot \sigma \cdot T_a^4 \cdot \left(1 + a \cdot N^b\right)$$

N	*fraction of cloud cover*
e, a	empirical coefficients

Longwave radiation

Longwave outgoing radiation (L_{out})

- L_{out} is emitted by the earth's surface.
- L_{out} is a function of the ***temperature*** and the ***emissivity.***
- L_{out} is constant (316 Wm^{-2}) under melting conditions (surface temperature = 0°C).
- L_{out} ***cannot exceed 316 [Wm^{-2}]*** over snow or ice.

- ***Snow and soil*** are commonly considered as ***grey bodies*** with emissivity close to 1, i.e. nearly black bodies.
- So the radiation emitted by the surface can be described with ***Stefan-Boltzmann law***.
- L_{out}, referring to the radiation ***emitted by*** and ***reflected from*** the ***surface***, can be calculated as:

$$\boxed{L_{out} = \varepsilon_s \cdot \sigma \cdot T_s^4 + (1 - \varepsilon_s) \cdot L_{in}}$$

1- ε_s reflectance

Emissivity (ε_s) of:

Snow:	0.98 -0.99
Ice:	~0.97
Soil:	0.95-0.97

6. Parameterization

Turbulent heat fluxes (Q_S and Q_L)

Sensible and **latent** heat fluxes are known as turbulent heat fluxes.

Sensible heat flux (Q_S)

Q_S is the heat energy transferred between the surface and air mass when a **difference in temperature** exits between them.

Q_S is a function of:

- ***temperature gradient*** between air temperature over the surface and the surface temperature
- ***wind speed***
- $Q_S = 0$ if the temperature gradient or the wind speed is zero.
- ***surface roughness***
- ***stability*** of the atmosphere

Latent heat flux (Q_L)

Q_L is the exchange of heat between the surface and air mass due to the ***change of phase*** of the water contained in the two media when there is a **difference in water vapour.**

Q_L is a function of:

- ***vapour pressure gradient*** between air and the surface
- ***wind speed***
- $Q_L = 0$ if the vapour pressure gradient or the wind speed is zero.
- ***surface roughness***
- ***stability*** of the atmosphere

Turbulent heat fluxes (Q_S and Q_L)

- They are driven by the ***temperature*** and ***moisture gradients*** between the air and surface, and by turbulence in the lower part of the atmosphere, i.e. by wind speed.
- The turbulent heat fluxes can be measured directly by ***eddy-correlation techniques***
  - ⇨ require sophisticated instrumentation
  - ⇨ continuous maintenance
  - ⇨ unsuitable for operational purposes
- Consequently such studies are rare and restricted to short periods.

⇩

Turbulent heat fluxes are often described by ***gradient flux relations***.

6. Parameterization

Turbulent heat fluxes

Gradient-Flux Relations

- The relations are based on the assumption of ***constant fluxes with height*** and ***horizontal homogeneous conditions***.
- Q_S and Q_L are ***proportional*** to the time-averaged gradients of ***potential temperature*** and ***specific humidity*** in the surface boundary layer.

$$\boxed{Q_S = \rho_a \cdot c_p \cdot K_S \cdot \frac{d\overline{\Theta}}{dz}}$$

$$\boxed{Q_L = \rho_a \cdot L_s \cdot K_L \cdot \frac{d\overline{q}}{dz}}$$

Q_S, Q_L	sensible and latent heat flux
ρ_a	density of the air
c_p	specific heat capacity of the air
L_S	latent heat of evaporation
K_S	eddy diffusivity for heat exchange
K_L	eddy diffusivity for vapour exchange
Θ	potential temperature
q	specific humidity

- K_S and K_L specify the ***effectiveness of the transfer process*** and depend on wind speed, surface roughness and atmospheric stability.

The profile method involves ***measurement*** of potential temperature, specific humidity and wind speed at preferably ***more than two levels*** above the surface.

→ **Bulk aerodynamic methods**

6. Parameterization

Turbulent heat fluxes

Bulk aerodynamic method

- Because detailed profile measurements are seldom available, one of ***numerous bulk aerodynamic methods*** is frequently applied for practical purposes.
- It exploits the fact that ***surface conditions*** of a melting surface are ***well defined*** (T=0°C, e=6.11 hPa)
- For the computation of Q_S and Q_L only ***one level of measurements*** is necessary.

$$Q_S = \rho_a \cdot c_p \cdot C_S \cdot u_a (T_a - T_s)$$

$$Q_L = \rho_a \cdot c_p \cdot C_L \cdot u_a (q_a - q_s)$$

Q_S	sensible heat flux [W m^{-2}]
ρ_a	density of the air [kg m^{-3}]
c_p	spec. heat capacity of air [J kg^{-1} K^{-1}]
C_S	bulk transfer coefficient for sensible heat [-]
u_a	wind speed at height z_a [m s^{-1}]
T_a	air temperature at height z_a [K]
T_s	temperature of snow surface [K]

Q_L	latent heat flux [W m^{-2}]
C_L	bulk transfer coefficient for vapour exchange[-]
q_a	specific humidity of air at height z_a [-]
q_s	specific humidity at the snow surface [-]

Kustas et al., 1994

- In practice it can be assumed $\boxed{C_S = C_L}$

6. Parameterization

Turbulent heat fluxes

Bulk aerodynamic method

- In the equation for the ***latent heat flux*** $(q_a - q_s)$ is often replaced by $(0.622/p_a)(e_a - e_0)$:

$$Q_L = \rho_a \cdot c_p \cdot C_L \cdot u_a (q_a - q_s)$$

$$Q_L = \rho_a \cdot c_p \cdot C_L \cdot u_a \cdot \frac{0.622}{p_a}(e_a - e_s)$$

Q_L	latent heat flux [W m^{-2}]
ρ_a	density of the air [kg m^{-3}]
c_p	spec. heat capacity of air [J kg^{-1} K^{-1}]
C_L	bulk transfer coefficient for vapour exchange[-]
u_a	wind speed at height z_a [m s^{-1}]
q_a	specific humidity of air at height z_a [-]
q_s	specific humidity at the snow/ice surface [-]
e_a	atmospheric vapour pressure at height z_a [Pa]
e_s	vapour pressure at the snow/ice surface [Pa]
p_a	atmospheric pressure [Pa]

6. Parameterization

Bulk aerodynamic method

- For ***neutral stability conditions*** in the boundary layer the bulk transfer coefficient is:

$$C_{Sn} = \frac{k^2}{\left[\ln\left(\frac{z_a}{z_0}\right)\right]^2}$$

C_{Sn}	bulk transfer coefficient for sensible heat for ***neutral stability*** atmosphere [-]
k	von Karman’s constant = 0.4
z_0	aerodynamic roughness length of snow or ice surface [m]

- The ***roughness length z_0*** for wind is defined as the ***height above the surface*** where the mean ***wind speed = 0***.
- It can be ***derived from detailed measurements*** of wind, temperature and humidity profiles.
- Over snow and ice z_0 can ***vary*** by ***several orders of magnitude***.
- Generally, ***z_0***-values of ***a few mm*** are often assumed.

Roughness length z_0:

snow: 0.004 - 70 [mm]
ice: 0.003 - 120 [mm]

6. Parameterization

Turbulent heat fluxes

Bulk aerodynamic method

- **Neutral stability** exists when the ***air*** and the ***surface*** have the ***same temperature***.
- Under ***neutral stability conditions*** (i.e. no buoyancy effects on turbulence) all turbulence is created by **forced convection** due to horizontal wind movement over a rough surface.
- **Free convection** refers to turbulence due to ***rising or sinking air*** caused by ***density differences*** in the air near the ground.
- In the atmosphere, turbulence is generally caused by a ***mixture of forced and free convection*** processes.
- Therefore, neutral stability conditions cannot be assumed.

A ***correction of the bulk transfer coefficient*** C_{Sn} is necessary.

6. Parameterization

Turbulent heat fluxes

Bulk aerodynamic method

- A variety of empirical expressions have been proposed to define the form of the stability functions.
- Due to its simplicity, the **Richardson number** is a commonly used ***stability criterion***.
- Its bulk form is defined as follows:

$$\boxed{Ri_b = \frac{g \cdot (T_a - T_s) \cdot z_a}{T_a \cdot u_a^2}}$$

Ri_b	bulk Richardson number [-]
g	acceleration due to gravity = 9.8 m s^{-2}
T_a	air temperature at height z_a [K]
T_s	temperature of snow surface [K]
u_a	wind speed at height z_a [m s^{-1}]

- **$Ri_b < 0$**: unstable stratification
- **$Ri_b > 0$**: stable stratification, which prevails over melting snow and ice surfaces.
- **$-0.01 < Ri_b < 0.01$**: neutral stability.
- ***Magnitudes*** of Ri_b (pos. or neg.) indicate the ***degree of instability or stability*** of the air.

6. Parameterization

Bulk aerodynamic method

- General stability correction equations given by Oke (1987):

Unstable conditions: $$\boxed{C_S = C_L = C_{Sn} \cdot (1 - 16Ri_b)^{0.75}}$$

Stable conditions: $$\boxed{C_S = C_L = C_{Sn} \cdot (1 - 5Ri_b)^{2}}$$

Ri_b	bulk Richardson number [-]
C_S	bulk transfer coefficient for sensible heat [-]
C_{Sn}	bulk transfer coefficient for sensible heat for ***neutral stability*** atmosphere [-]
C_L	bulk transfer coefficient for vapour exchange[-]

6. Parameterization

Turbulent heat fluxes

Bulk aerodynamic method

Atmospheric stability and snowpack convection conditions

Atmospheric Stability	Meteorological Conditions	Occurrence	Bulk Richardson Number	C_h/C_{hn}
Unstable-free convection	$T_s >> T_a$, u_a relatively low	Rare over snow	[illegible]	$>>1$
Unstable-mixed convection	$T_s > T_a$, u_a relatively high	Occasional; windy, cold winter accumulation season, night time	<-0.01	>1
Neutral-forced convection	$T_s \sim T_a$, u_a relatively high	Common; windy, cool periods, melt initiation	-0.01 to $+0.01$	~ 1
Stable-damped forced convection	$T_s < T_a$, u_a relatively high	Common; windy, warm melt periods	$>+0.01$	<1
Stable-fully damped forced convection	$T_s << T_a$, u_a relatively low	Common; calm, warm melting periods; nights with fog and cold air drainage	$\rightarrow$critical value $\sim +0.2$ to $+0.4$	$\rightarrow 0$

Als Backhand-Folie gedacht

DeWalle, D.R. and Rango, A.: 2008

Wind-Index Method

- The wind-index methods are a further ***simplification*** compared to the bulk aerodynamic methods.
- Wind-index methods use an ***empirical wind function*** to calculate the turbulent energy fluxes.
- They are often used when calculating melt rates using data from operational weather stations.
- The following examples were developed by ***Anderson, E. A.*** and ***Braun, L.***

Sensible heat

$$\boxed{M_S = (C1 + C2 \cdot u) \cdot T_a / S}$$

M_S	melt due to sensible heat [mm/time interval]
$C1$	wind function coefficient independent of wind speed [mm d^{-1} C^{-1}]
$C2$	wind function coefficient associated with wind [mm d^{-1} C^{-1} (m s^{-1}) $^{-1}$]
u	wind speed [m s^{-1}]
T_a	air temperature [°C]
S	number of time steps during one day

Wind-Index Method

Latent heat

$$\boxed{M_L = (C1 + C2 \cdot u) \cdot (E(T_a) - 6.11) / \gamma \cdot S}$$

$$\boxed{M_L = (C1 + C2 \cdot u) \cdot (e_a - 6.11) / \gamma \cdot S}$$

M_L	melt due to latent heat [mm/time interval]
$C1$	wind function coefficient independent of wind speed [mm d^{-1} C^{-1}]
$C2$	wind function coefficient associated with wind [mm d^{-1} C^{-1} (m s^{-1}) $^{-1}$]
u	wind speed [m s^{-1}]
T_a	air temperature [°C]
e_a	atmospheric vapour pressure [hPa]
E	saturated vapour pressure at temperature T_a [hPa]
Υ	psychrometric constant
S	number of time steps during one day

6. Parameterization

'Cold Content' of Snow and Ice (Q_I)

- Before surface melting can occur, the temperature of the snow/ice surface must be raised to 0°C.
- The energy required to heat a cold snow pack or ice to 0°C is called '**cold content'.**
- Changes in the cold content lead to ***changes*** in the internal energy storage (***Q_I***)

The '**cold content'** ***C*** of a column of snow/ice of depth *z* below the surface is given by:

$$\boxed{C = -\int_0^z \rho(z) \cdot c_p \cdot T(z) \cdot dz}$$

C	'cold content'
ρ(z)	density at depth z [kg m^{-3}]
c_p	specific heat capacity of snow/ice [J kg^{-1} K^{-1}]
T(z)	temperature at depth z [°C]
z	depth [m]

Specific heat capacity:

snow:	2009 [J kg^{-1} K^{-1}]
ice:	2097 [J kg^{-1} K^{-1}]
water:	4180 [J kg^{-1} K^{-1}]
air:	1005 [J kg^{-1} K^{-1}]

Energy required to warm up 1 kg by 1 K.

6. Parameterization

Heating of Snow and Ice

- ***Melt water*** produced at the surface, or ***rain***, **percolates** down through the snow pack.
- Rainfall onto a subzero snowpack would initially give up its ***sensible heat***.
- If that was not sufficient to warm the snow to 0 °C, the rain water would **refreeze** an release the latent heat of fusion.
- ***Re-freezing of 1 g water*** releases the energy to ***heats 165 g snow by 1*** °C.

Latent heat of fusion: 0.334 MJ kg^{-1}	⟷	Specific heat capacity 0.00201 J kg^{-1} K^{-1}

- The 'cold content' is equal to 0 if the snow is at 0°C at all depths. The snow or ice is then referred to as **temperate**.
- The 'cold content' of a winter snow cover or the surface ice layers can be an important **retention** component .
- The 'cold content' can contribute significantly to the delay between surface melt and melt derived runoff.

Liquid Water Content of Snow

Table 1. Hand test for the qualitative estimation of liquid water content (mWC) and the approximate range of liquid water content (). The detailed description is taken from the International Classification of Seasonal Snow on the Ground (Fierz et al., 2009, p. 8). This classification is also used in Swiss observational guidelines (WSL, 2008). Half index classes may also be used. ts – snow temperature.

Wetness Content	Index (mWC)	Description	θ [vol. %]
Dry	1	$t_s \leq 0.0$ °C. Disaggregated snow grains have little tendency to adhere to each other when pressed together.	0
Moist	2	$t_s = 0.0$ °C. The water is not visible, even at 10× magnification. When lightly crushed, the snow has a tendency to stick together.	0–3
Wet	3	$t_s = 0.0$ °C. The water can be recognized at 10× magnification by its meniscus between adjacent snow grains, but water cannot be pressed out by moderately squeezing the snow in the hands.	3–8
Very Wet	4	$t_s = 0.0$ °C. The water can be pressed out by moderately squeezing the snow in the hands, but an appreciable amount of air is confined within the pores.	8–15
Soaked	5	$t_s = 0.0$ °C. The snow is soaked with water and contains a volume fraction of air from 20 to 40 %.	>15

6. Parameterization

Energy from rain

Energy from Rain (Q_R)

Rainfall can influence the energy budget of a snowpack in three ways:

1. Addition of ***sensible heat*** due to heat added by relatively warm rain.
2. Release of ***latent heat of fusion***, if rain freezes on a sub-zero snowpack.
3. C***ondensation*** at the surface due to high humidity associated with rainy weather.

1. The **sensible heat input by rain** can be computed as:

$$\boxed{Q_R = \rho_w c_w R(T_r - T_s)}$$

Q_R	Sensible heat of rain [W/m²]
ρ_w	density of water [kg/m³]
c_w	specific heat capacity of water [J kg⁻¹ K⁻¹]
R	rainfall intensity [m/s]
T_r	temperature of rain (air) [°C]
T_s	surface temperature (snow/ice) [°C]

- The sensible heat of rain is ***generally unimportant*** to other energy balance comp.
- It contributes usually not more then ***a few percents*** of the energy for melting.
- ***Exception***: heavy, prolonged and warm rainfall, as in areas with advection of air originating over warm oceans.

Indirect effects of rain:

- ➢ increasing liquid water content of snow
- ➢ albedo of snow decreases
- ➢ increasing absorption of solar radiation
- ➢ more energy for melt

6. Parameterization

Heat from ground

Ground Heat Flux (Q_G)

- Ground heat conduction to the base of the snowpack generally represents a ***very minor energy source*** for melt because:
  - ***soil*** is in general a ***poor conductor*** of heat
  - ***soil temperatures*** are often ***low*** beneath a snowpack
- During the accumulation season ground heat can contribute to gradual ***ripening*** and ***slow melting of basal snow layers***.
- This contribution diminishes into spring when the snowpack becomes warmer.

The heat conduction flux Q_G can be computed as:

$$\boxed{Q_G = k\ \frac{dT}{dz} \cong k \frac{(T_{z1} - T_0)}{(z_1 - z_0)}}$$

Thermal conductivity:

Quartz:	8.80 [W m^{-1} K^{-1}]
Clay minerals:	2.92 [W m^{-1} K^{-1}]
Organic matter:	0.25 [W m^{-1} K^{-1}]
Water [0°C]:	0.56 [W m^{-1} K^{-1}]
Ice [0°C]:	2.24 [W m^{-1} K^{-1}]
Air:	0.025 [W m^{-1} K^{-1}]
Soil:	~ 0.2–2 [W m^{-1} K^{-1}]

Q_G	ground heat flux [W m^{-2}]
k	thermal conductivity of soil [W m^{-1} K^{-1}]
$\frac{dT}{dz}$	temperature gradient in the soil [K m^{-1}]
T_{z1}	soil temperature at depth z_1 [K]
T_0	soil temperature at base of snowpack z_0 [K]

Melt Energy (Q_M)

- At a surface temperature of 0 °C, any surplus of energy is assumed to be used for melting.
- In snow and glacier melt models, melt energy (Q_M) is generally solved as a residual of the energy balance equation:

$$\boxed{Q_M = Q_{NR} + Q_S + Q_L + Q_P + Q_G}$$

- The energy available for melt Q_M can be converted to the **melt rate M**:

$$\boxed{M = \frac{Q_M}{\rho_w L_f}}$$

Q_M = 100 W m^{-2} ➡ M = 26 mm/day

M	Melt rate of ice or snow [m s^{-1}]
Q_M	heat used for melt [W m^{-2}]
ρ_w	density of water [kg m^{-3}]
L_f	latent heat of fusion (334 kJ kg^{-1} at 0°C)

Contents

1. Introduction
2. Blackbody radiation
3. Solar radiation
4. Measurement of energy balance components
5. Special Characteristics of Snow and Ice
6. Parameterization of energy balance components
7. Typical values and relative importance of energy balance components

7. Typical values and relative importance

Relative importance of energy balance components

Point energy balance studies on Alpine valley glaciers

- The values of the energy balance components are in % of total energy source or sink.
- Different reference periods!

Location	Lat.	m a.s.l.	Q_{NR} [%]	Q_S [%]	Q_L [%]	Q_G [%]	Q_M [%]	Reference
Aletschglacier	46°26' N	3366	92	8	-6		-94	Röthlisberger & Lang, 1987
Vernagtferner	46°52' N	2970	84	14	2	0	-100	Hoinkes, 1955
Antizana Gl. 15	0°28' S	4890	80	20	-26		-74	Favier et al., 2004b
Morteratsch Gl.	46°24' N	2100	80	16	4		-100	Oerlemans, 2000
Pasterze Glacier	47°5' N	2205	74	21	5		-100	van den Broeke, 1997
Aletschglacier	46°26' N	2220	71	21	8		-100	Röthlisberger & Lang, 1987
Storglaciären	67°54' N	1370	65	30	5	-3	97	Hock & Holgren, 1996
Hodges Glacier	54°16' S	460	54	46	-3		-97	Hogg et al., 1982

7. Typical values and relative importance

Relative importance of energy balance components

- The relative importance of the energy balance components ***depends strongly on weather conditions***, e.g. whether there are ***clear sky*** or ***overcast*** conditions.

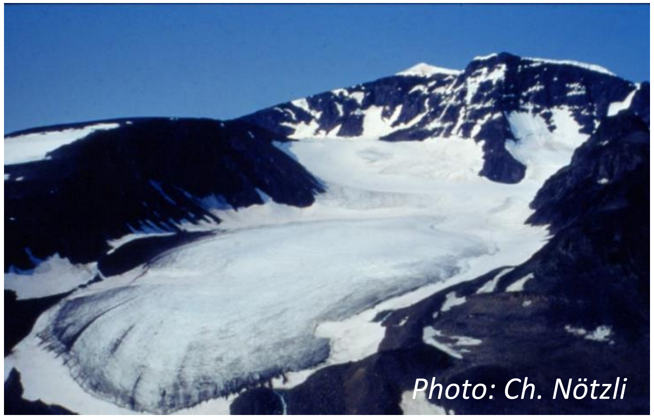


Photo: Ch. Nötzli

- In general, most of the energy used for melt is supplied by **net radiation** followed by the **sensible heat** flux.
- **Latent heat** often plays only a ***minor role***.
- As regards net radiation, **incoming longwave radiation** is by far the largest source of energy for melt.

- The ***importance of net radiation*** tends to ***increase with altitude*** due to:
  - decrease of air temperature with altitude
  - decrease of vapour pressure with altitude

  → ***decrease*** of ***turbulent fluxes***

- Generally, the energy from **rain** is relatively ***small***.
- The contribution of the **ground heat flux** to melt is almost always ***negligible***
- A very ***intense snowmelt*** often occurs with ***large turbulent heat transfer***.

7. Typical values and relative importance

Relative importance of turbulent heat fluxes

- The turbulent fluxes of sensible and latent heat are ***generally small*** when averaged over periods of weeks and months and compared to the net radiation flux.

- Over ***short time intervals*** of hours and days turbulent heat fluxes can exceed net radiation.
- On ***cloudy*** and rainy days turbulent fluxes can dominate the energy balance.
- ***Highest melt rates*** often coincide with ***high values of the turbulent fluxes***.

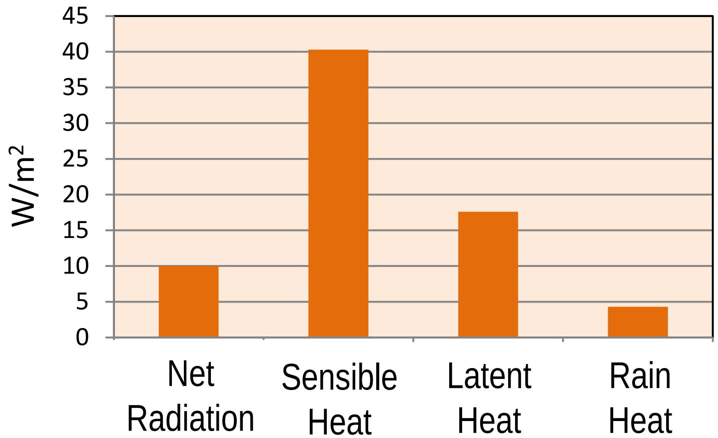
- ***Sublimation*** can be important at ***high altitudes*** and ***high latitudes*** (e.g. blue-ice areas, Antarctica)
- A result of high sublimation are the so called ***Snow Penitentes***.

7. Typical values and relative importance

Relative importance of energy balance components

Example of a combined rain-snowmelt event on January 26, 1976

Rietholzbach, Switzerland (Source: Lang, H.,1986)



	Net Rad.	Sensible Heat	Latent Heat	Rain Heat	Melt Energy
[W m^{-2}]	10.0	40.3	17.6	4.3	72.2
[%]	14	56	24	6	100
[mm d^{-1}]				21.5	18.7

- ***Sensible heat*** was the ***dominating*** heat source.
- ***High air humidity***, due to ***rainfall*** and ***high air temperatures***, produced condensation conditions with latent heat contribution of 24% to the melt process.
- Incoming ***solar radiation*** was ***strongly reduced*** because of full cloud coverage.
- ***Longwave net radiation*** was ***slightly positive*** -> total net radiation 14%.
- Energy provided from ***rain*** was of ***minor importance*** (6 %).

Energy Balance over snow and ice

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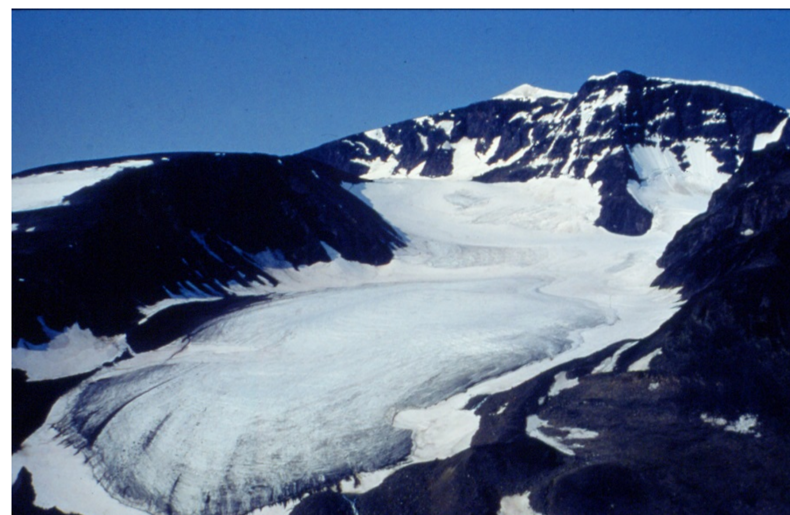
Notizen

METEOROLOGIC DATA REQUIREMENTS FOR SNOW SIMULATION

Meteorologic Data	Energy Balance	Temperature Index
Precipitation	Required	Required
Air Temperature	Required	Required
Solar Radiation	Required	Not Used
Dewpoint	Required	Optional
Wind Velocity	Required	Not Used
Cloud Cover	Optional	Not Used



7. Typical values and relative importance

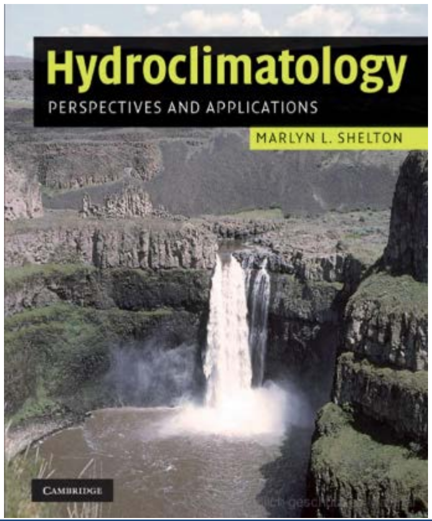
Table 2 Point energy-balance studies on Alpine valley glaciers. Net radiation Q_N, sensible heat flux Q_H, latent heat flux Q_L, ice heat flux Q_G and the energy for melt Q_M (here defined as negative) are given in W m^{-2}. Values in brackets are in % of total energy source or sink. The energy balance does not necessarily balance if Q_M is obtained from ablation measurements instead of from closing the energy balance

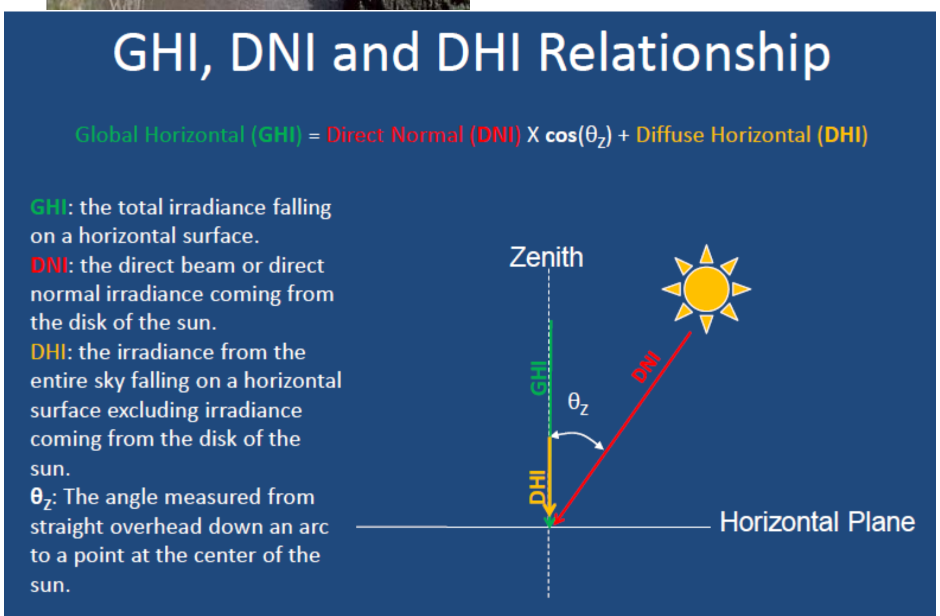
Location, m a.s.l., surface type	Period (d = days)	Q_N	Q_H	Q_L	Q_G	Q_M	Reference
Vernagtferner 2970 m, ice	45 d in Aug + Sep 1950–53	143 (84)	23 (14)	4 (2)	0 (0)	−170 (−100)	Hoinkes, 1955
Kesselwandferner 3240 m, snow	20 d in 1958	43 (67)	21 (33)	−1 (−2)		−64 (−98)	Ambach and Hoinkes, 1963
Blue Glacier 2050 m	12.7.–20.8.1958	85 (63)	50 (37)	−3 (−2)		−132 (−98)	La Chapelle, 1959
Aletschglacier 2220 m, ice	2.–27.8.1965	129 (71)	38 (21)	14 (8)		−181 (−10[illegible])	Röthlisberger and Lang, 1987
Aletschglacier 3366 m, snow	3.–19.8.1973	44 (92)	4 (8)	−3 (−6)		[illegible] (−94)	Röthlisberger and Lang, 1987
Worthington Glacier Alaska, ice	16.7.–1.8.1967	127 (51)	68 (29)	47 (20)		−224 (−100)	Streten and Wendler, 1968
Peytoglacier 2510 m	14 d in July 1970	80 (44)	87 (48)	[illegible]		−181 (−100)	Föhn, 1973
Hodges Glacier South Georgia, 460 m	1.11.1973–4.4.1974	47 (54)	42 ([illegible])	−3 (−3)		−86 (−97)	Hogg *et al.*, 1982
St Sorlin Glacier 2700 m	11 d in summer	32 (57)	24 (43)	−4 (−7)		−53 (−93)	Martin, 1975
Hintereisferner ice	10 d in 1986	191 (90)	22 (10)	−4 (−2)		−209 (−98)	Greuell and Oerlemans, 1987
Ivory Glacier, New Zealand, 1500 m	53 d in Jan–Feb 1972/73	76 (52)	44 (30)	23 (16)	*	−147 (−100)	Hay and Fitzharris, 1988
Storglaciären 1370 m, ice	[illegible]7.8.1994	73 (66)	33 (30)	5 (5)	−3 (−3)	−122 (−97)	Hock and Holmgren, 1996
Paterze glacier 2205 m, ice	24.6.–9.9.1994	180 (74)	51 (21)	11 (5)		−242 (−100)	van den Broeke, 1997
Zongo Glacier 5150 m, ice/snow	9/1996–8/1997	17 (65)	6 (23)	−17 (−65)	3 (12)	−9 (−35)	Wagnon *et al.*, 1999
Morteratschgletscher** 2100 m, ice/snow	1.10.1995–30.9.1998	152 (80)	31 (16)	8 (4)		−191 (−100)	Oerlemans, 2000
Koryto Glacier, Kamchatka, 840 m, snow	10.8.–8.9.2000	43 (33)	59 (44)	31 (33)		−133 (−100)	Konya *et al.*, 2004

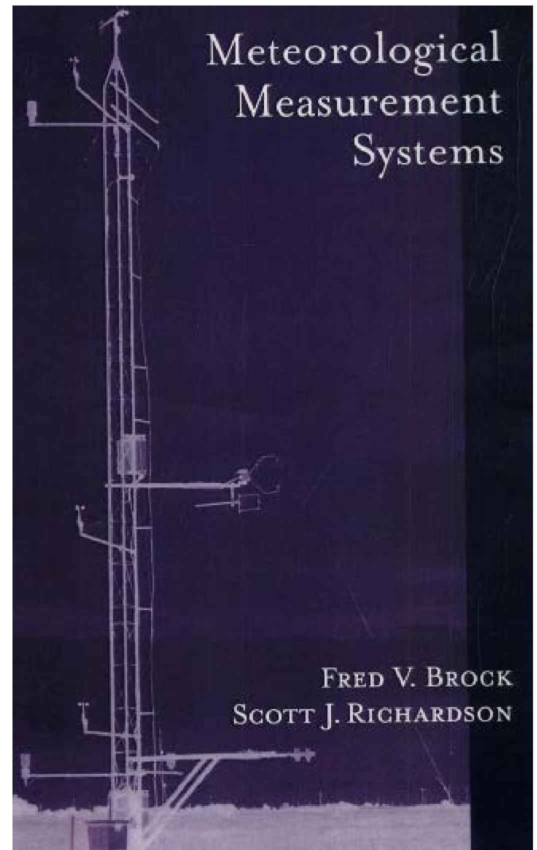
*Rain supplied 4 W m^{-2} (2%).
**Only when melting occurred.

Hock, 2005

http://www.meteorologynetwork.com/dokuwiki/doku.php?id=themen:atmosphaere:geraete:strahlungsmessung







Energy Balance

Links for **radiation measurement**

http://almashriq.hiof.no/lebanon/600/610/614/solar-water/unesco/21-23.html

http://www.thiesclima.com/radiation_glossary.html

http://rredc.nrel.gov/solar/pubs/bluebook/appendix.html

http://www.seco.cpa.state.tx.us/publications/renewenergy/solarenergy.php

http://solardat.uoregon.edu/SolarRadiationBasics.html
http://www.bom.gov.au/climate/austmaps/solar-radiation-glossary.shtml

http://wiki.naturalfrequency.com/wiki/SolarRadiation/Components

The sun's position (latitude, hour angle, zenith angle, etc
http://www.powerfromthesun.net/Book/chapter03/chapter03.html

http://almashriq.hiof.no/lebanon/600/610/614/solar-water/unesco/21-23.html
http://ww2010.atmos.uiuc.edu/(Gl)/guides/mtr/opt/mch/sct.rxml
http://personal.cityu.edu.hk/~bsapplec/solar2.htm
http://www.geog.ucsb.edu/~joel/g110_w08/lecture_notes/radiation_atmosphere/radiation_atmosphere.html

http://ww2010.atmos.uiuc.edu/(Gl)/guides/mtr/opt/mch/sct.rxml

http://www.wetterochs.de/wetter/feuchte.html