Universität Hamburg
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MIN-Fakultät
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Arbeitsbereich SAV/BV (KOGS)

IP2: Image Processing in Remote Sensing
5. The Atmosphere of the Earth

Summer Semester 2014

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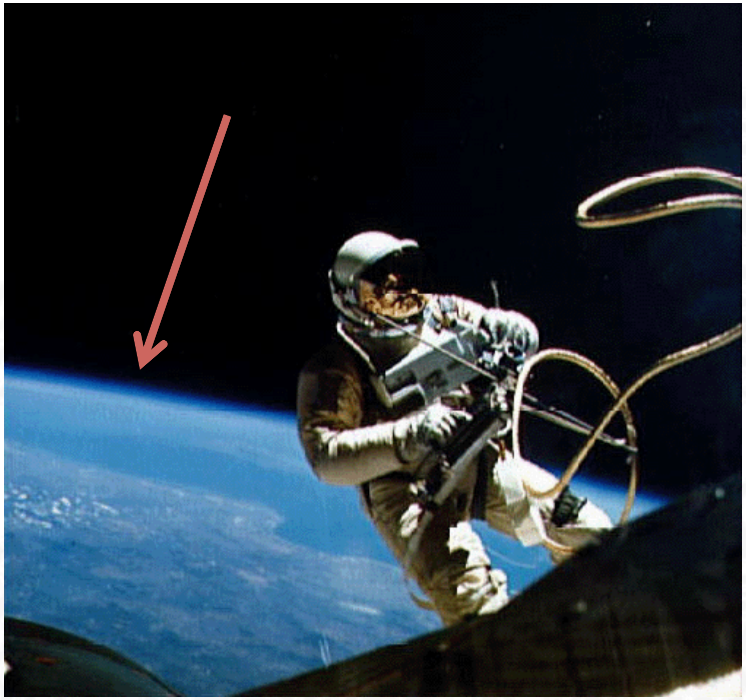
Agenda

- Structure
  - Pressure
  - Temperature
  - Chemical
- Optical Properties
  - Atmospheric Opacity
  - Refraction
  - Global radiation
  - Acquisition parameters

The Atmosphere: A Volatile Thin Layer

- Necessary for life at earth (and maybe elsewhere)
- Influences on EM radiation:
  - Refraction
  - Absorption
  - Scattering
- Important properties:
  - Density, Pressure, Temperature
  - Chemical composition
  - Ionization and Magnetic Properties



Apollo in Orbit

Air Pressure (1)

- The atmosphere in a hydrostatic balance:
  - Gravity acts on each volume element (↓)
  - Pressure gradient force acts in the opposite direction (↑)

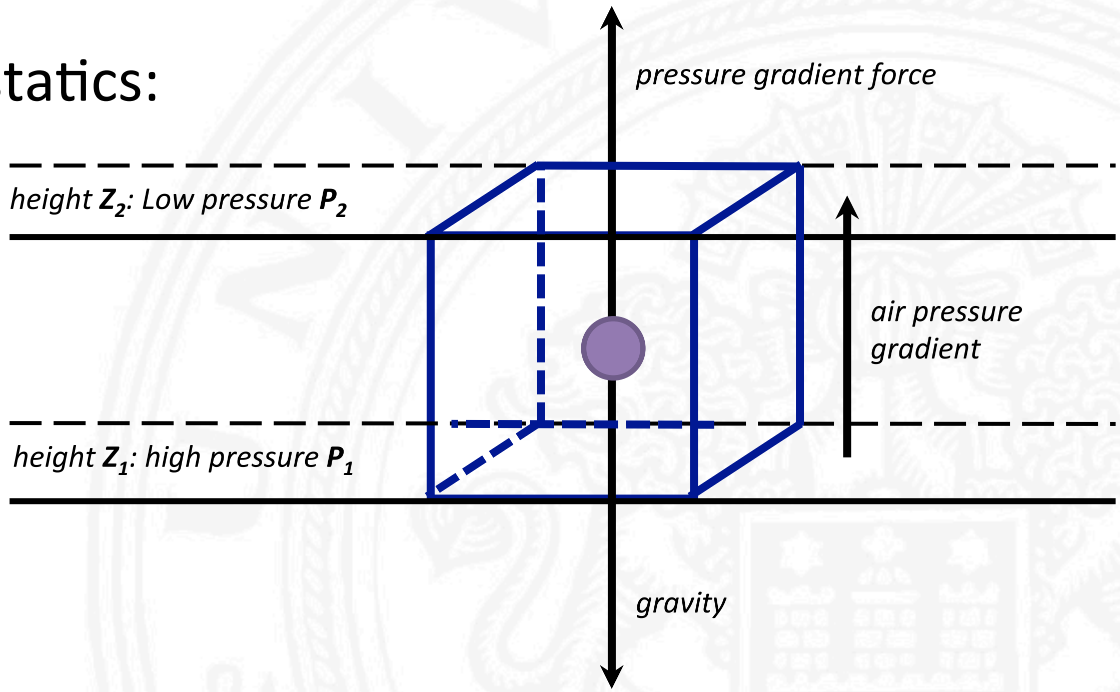
- Euler's Law of Hydrostatics:

$$\Delta P = P_1 - P_2$$
$$= -(Z_1 - Z_2)\rho g$$

with:

g gravity

ρ density



Air Pressure (2)

- The pressure difference $P_1 - P_2$ is proportional to:
  - the vertical distance $Z_1 - Z_2$ of both pressure fields
  - the air density ρ between both levels
  - the gravity g
- Gravity and pressure gradient force act perpendicular to the earth's surface
  - Atmosphere consists of vertically ordered level of equal pressure
- Mass distribution
  - 90% of the mass of the atmosphere → lower 16 km
    Air pressure at 16 km → 10% of pressure on ground elevation
  - 99% of the mass in the lower 30 km!

Temperature and Pressure

- The larger the pressure difference the larger the elevation difference (at constant density)
- Gas Laws state that the temperature is inverse proportional to the density (at equal pressure)
  - The larger the density or the lower the temperature, the lower the vertical difference between the pressure levels
  - The lower the density or the larger the temperature, the larger the vertical difference between the pressure levels

Barometric Scale Factor

Barometrische Höhenstufe

- The elevation difference, on which the air pressure changes for 1 hPa is called Barometric scale factor:
  - Basis: Euler's Law of Hydrostatics:

    $$\Delta P = -\mathrm{d}z\rho g$$

  - At sea level ($\rho = 0.12923 \text{ kg m}^{-3}$):

    $\Delta P = 1 \text{ hPa} \rightarrow \Delta Z = 7.89 \text{ m}$

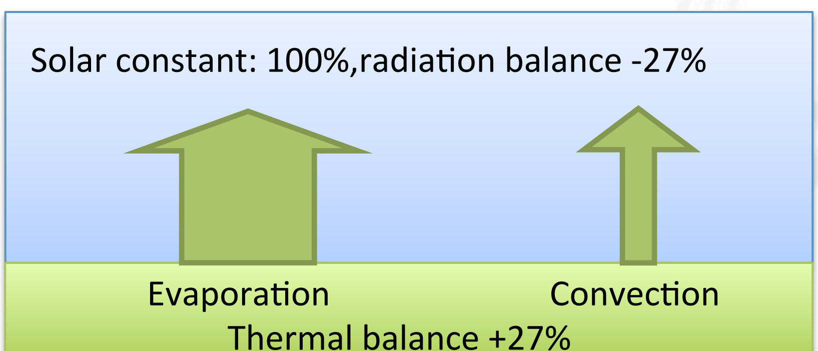
  - At 6 km elevation ($\rho = 0.06462 \text{ kg m}^{-3}$):

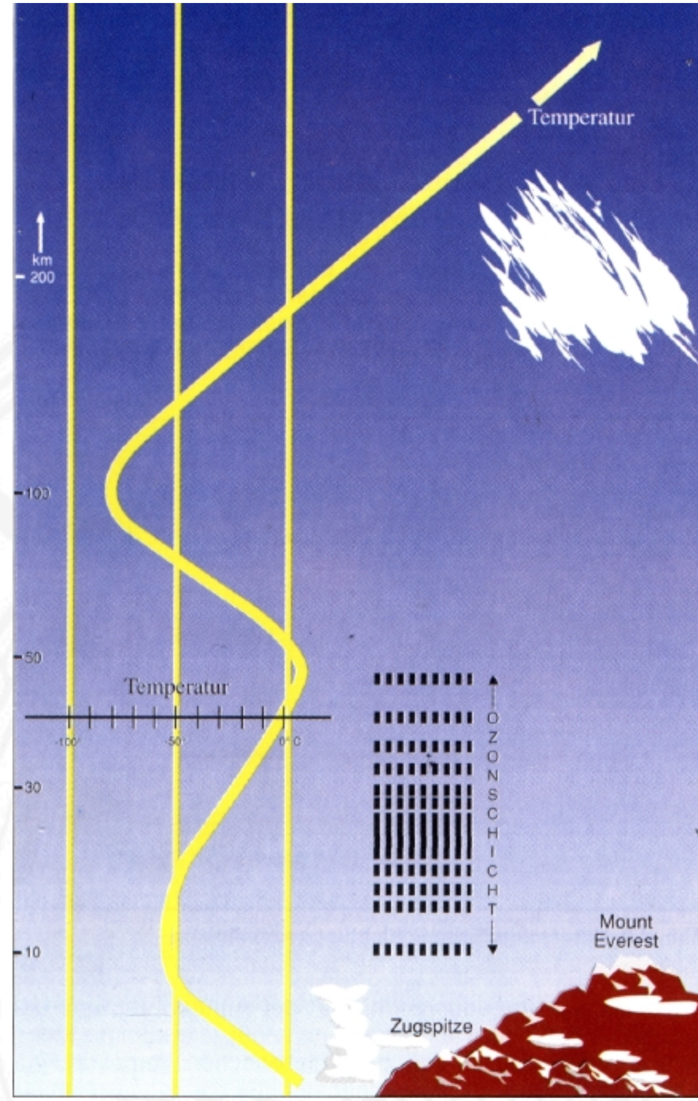
    $\Delta P = 1 \text{ hPa} \rightarrow \Delta Z = 15.78 \text{ m}$

- Observations:
  - The barometric scale factor depends on the the density
  - The air pressure logarithmically diminishes

Vertical Structure of the Atmosphere: Temperature

- Non-linear temperature distribution
- Caused by different energy sources
  - Irradiation of the sun
  - Thermal radiation from earth's surface
  - Convection
  - Evaporation
    Verdunstung
- Thermal balance



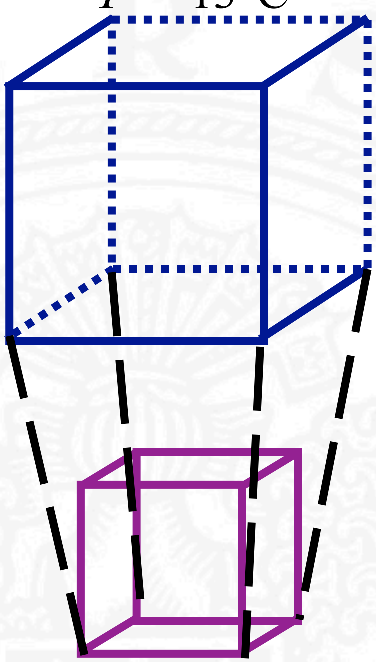


Thermal Convection

- Air layers close to the ground are heated and thus become lighter.
- Due to the differences of the density, the (heated) air is elevated.
- Du to the loss of pressure in higher elevations, the (heated) air reaches areas of lower pressure and thus expands.
- This expansion yields to a loss of thermal energy of the (heated) air.
- If there is no energy exchange between the environment and the cell, it is called an adiabatic process.

Adiabatic Height Scale

- The adiabatic cooling rate on decent/ascent is nearly 1°C per 100 m height difference.
- As soon as the temperature is lower than the dew-point, the enclosed moisture begins to condensate.
- The emitted condensation warmth acts opposite to the adiabatic cooling. The cooling rate decreases to approx. 0.65°C per 100 m.



Dew-point: *Taupunkt*	The temperature where the (enclosed) water vapor starts to condensate due to saturation

Temperature Layering of the Atmosphere

0 km	**Troposphere:**	Lowest layer with a temperature reduction of (at mean) 0.65°C/100 m. Highest amount of water vapor
15 km	**Tropopause:**	Upper boundary of the Troposphere (at poles: 8 km)
	Stratosphere:	
	Isotherm:	Temperatures at a mean of -56.5°C
	Warm:	Temperature increases up to 0°C (Ozone absorption)
50 km	**Stratopause:**	Upper boundary of the Stratosphere
	Mesosphere:	Temperature diminishes up to -80°C
80 km	**Mesopause:**	Upper boundary of the Mesosphere
	Thermosphere:	Warming through O_2 absorption. Low density, high temperatures (up to 100°C)
500 km		

Chemical Composition of the Atmosphere

0 km

Homosphere:

Gases inside this layer are highly mixed, due to strong air movements by convection and winds. The mixture ratio of all gases is approximately constant.

30 km O_2 is dissociated, but recombined immediately to O_3 or O_2.

80 km Solar UV radiation dissociates CO_2 to C und O_2.

Heterosphere:

100 km O remains atomic.

110 km Diffusion dominates the gas mixture. Molecules are separated by their weight → diffusion balance.

700 km Mean molecular weight diminished to 16.

The Exosphere (Z>1000 km)

- Inside the Exosphere, uncharged particles can reach the escape speed (from earth's gravity)
  - High temperatures give the particles high velocities.
  - (Very) small density, only few particles give long free space trajectories for each particle.
  - Lighter gases escape more often to space than heavier ones.

Ionosphere (Z > 80km)

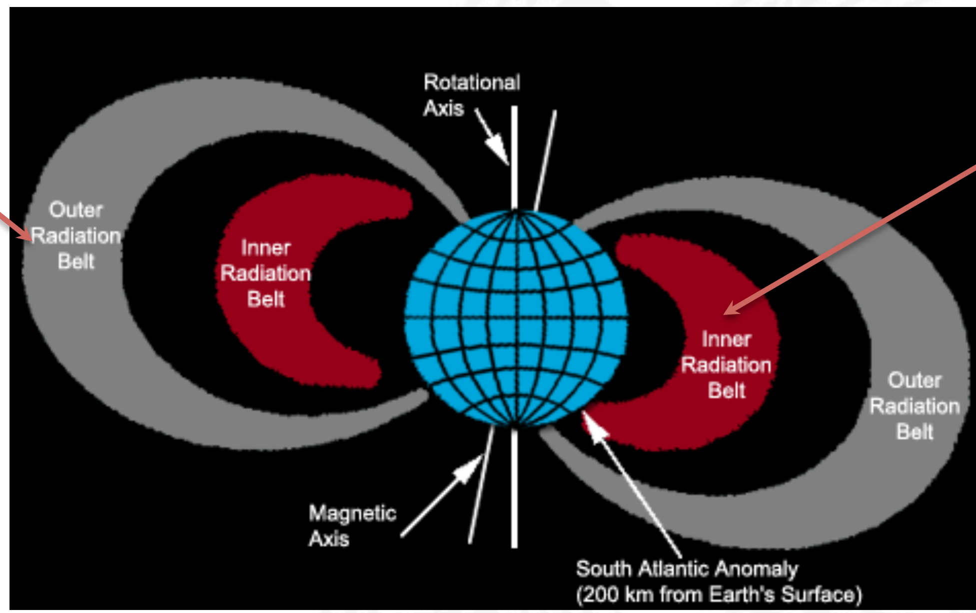
- Ionization has different sources:
  – highly energetic UV radiation and X-Rays
  – Particle radiation: mainly electrons and protons
- Ionization has effects on wireless communications:
  – Radio waves are slowed down by the Ionosphere. This needs to be compensated e.g. by GPS-measurements, since the are based on time of flight!
  – The atmosphere is completely opaque for radio waves of frequencies < 9 MHz. This allows short wave radio around the complete earth, since the (short) waves are reflected back to earth at the Ionosphere.

Magnetosphere (1 000 – 100 000 km)

- Between 1 000 and 100 000 km altitude, the magnetic field of the earth influences the movement of charged particles.
- Defines the atmospheric boundary
- Contains both Van-Allen radiation belts:

electrons

protons



Different Characterizations of the Atmosphere

<table>
<tr><th>Temperature</th><th>Chemical composition</th><th>Ionization</th><th>Escape movement</th></tr>
<tr><td colspan="4">Interplanetary space</td></tr>
<tr><td>Thermosphere</td><td rowspan="2">Heterosphere</td><td>Protonsphere</td><td>Exosphere</td></tr>
<tr><td>Mesosphere</td><td>Ionosphere</td><td rowspan="3"></td></tr>
<tr><td>Stratosphere</td><td rowspan="2">Homosphere</td><td rowspan="2">Neutrosphere</td></tr>
<tr><td>Troposphere</td></tr>
<tr><td colspan="4">Earth surface</td></tr>
</table>

Optical Properties of the Atmosphere

- Atmospheric Extinction (*Trübung*)
  - Scattering
    - Molecules: Rayleigh
    - Aerosols: Mie
    - Dust & drops: non-selective
  - Absorption
- Refraction
- Global radiation
- Upward sky radiation
  Luftlicht
- Acquisition Parameters:
  - Max. visible distance
  - Contrast

Scattering of EM Waves
Wavelength Selectivity

The kind and behavior of scattering depends on the size of the scatterers s:

1. **s ≫ λ: non-selective:**
   The radiation of all wavelengths is similar. Thus, the scattering on large particles (like dust) is non-selective
2. **s ≪ λ: Rayleigh scattering:**
   Resonance effects between radiation and particles result in a wavelength depending scattering. Thus, the scattering at molecules () is selective
3. **s ≈ λ: Mie scattering:**
   Only weak selectivity, like haze.

Scattering of EM Waves
Rayleigh Scattering (1)

Example:

Rayleigh scattering on air molecules (radiation source at azimuth).
The scattering coefficient k is:

$$k_\lambda = \frac{5 \cdot 32\pi^3 (n-1)^2 H_0}{3 \ln(100) \lambda^4 N} \quad \rightarrow \quad k_\lambda \propto \lambda^{-4}$$

with:

- n refraction index of the air
- H_0 Barometric scale
- N density of the atmosphere

The scattering of blue light is very high compared to red light.

- If a spectrum of light reaches the air, blue light is scattered most often at air molecules
- At sunset, only red light reaches the earth, the environment of the sun lights red

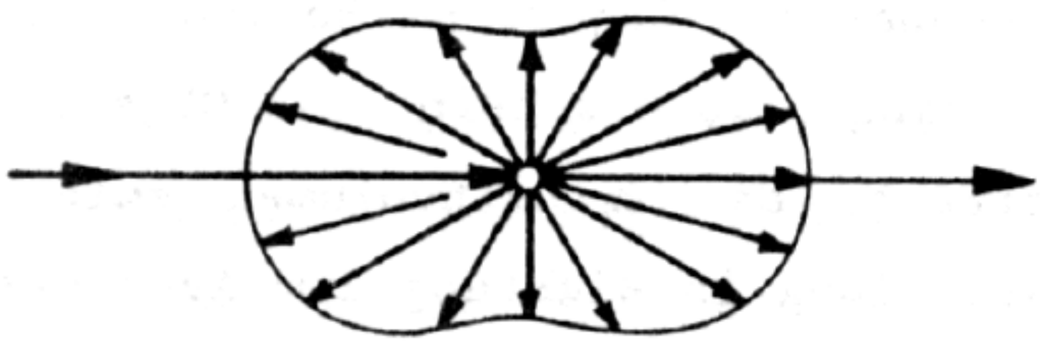
Scattering of EM Waves
Rayleigh Scattering (2)

Direction dependency: The Rayleigh scattering does not uniformly scatter in each direction!

- Parallel to the trajectory ,the scattering is twice as large as perpendicular to this direction

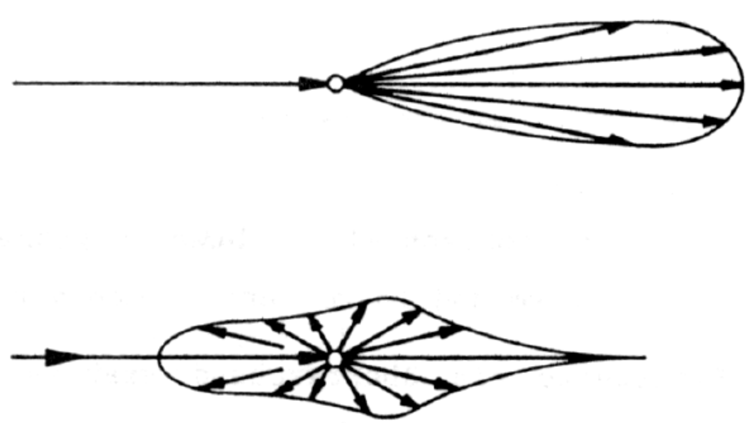
→ The sky is comparably dark in an 90° angle to the sun.

→ In a small environment around the sun, sky is much brighter



Scattering of EM Waves
Mie Scattering (1)

- The scattering at aerosols and small dust particles is not as wavelength depending as the molecular scattering.
- But: dust may scatter 1000x higher in forward direction than side- or backwards
- The scattering coefficient is given by:
  $k_\lambda = \beta \cdot \lambda^{-a}$ with $1 < a < 4$ (mean: a=1.3)
- The extinction coefficient β as well as the exponent a are weather dependent (due to the size and count of different particles)



Scattering of EM Waves
Mie Scattering (2)

- Mie/Dust scattering results in:
  - white scattering since all wavelengths are similarly scattered
  - an additive, bright veil on the images
- Special case: Aureole (right image)
  due to strong forward scattering of the radiation

Scattering of EM Waves
Non-selective Scattering

- If particles are much larger than the wavelength, non-selective scattering occurs:
- Independent of the wavelength:

$$k_\lambda \propto \lambda^0$$

- First Example for non-selective scattering on dust particles:

Scattering of EM Waves
Non-selective Scattering

- Second example: Sandstorm in Tehera (Iran) @ 06/02/2014:

https://www.youtube.com/watch?v=0nlqrk6IMx0

Absorption

- Caused by molecules:
  - Electron absorption
  - Vibration
  - Rotation
- Results in spectral absorption lines (right, from Rees 1990)

Visible light & Infrared	
Molecule	**Wavelength [µm]**
H_2O	0.9,1.1,1.4,1.9, 2.7, ≈6
O_2	1
CO_2	2.7,4.3, ≈14
N_2O	4.6,7.7
O_3	10
Microwaves	
Molecule	**Frequency [GHz]**
H_2O	22.235,183.3
O_2	≈60,118.75

Example:

Ozone has a narrow absorption band (Cappius band) at green range.

→ Sky looks not green, even if the sun is low over the horizon and there is nearly blue light available for scattering!

Atmospheric Extinction

Atmosphärische Trübung

- The reduction of radiation w.r.t scattering has the same (exponential) form than the reduction by absorption.

Scattering: $I_\lambda = I_{0_\lambda} e^{-k_\lambda \frac{p}{p_n} m_r}$ **Absorption:** $I_\lambda = I_{0_\lambda} e^{-\mu_\lambda \frac{p}{p_n} m_r}$

- Atmospheric Extinction combines both influences:

$I_\lambda = I_{0_\lambda} e^{-ext_\lambda \frac{p}{p_n} m_r}$ with: $ext_\lambda = k_\lambda + \mu_\lambda$

- The extinction yields to an exponential decrease of radiation (depending on the relative air mass). Thus it is often logarithmically expressed (in decibel [dB]):

$$E = 10 \cdot \log_{10}\left(\frac{I_1}{I_2}\right)$$

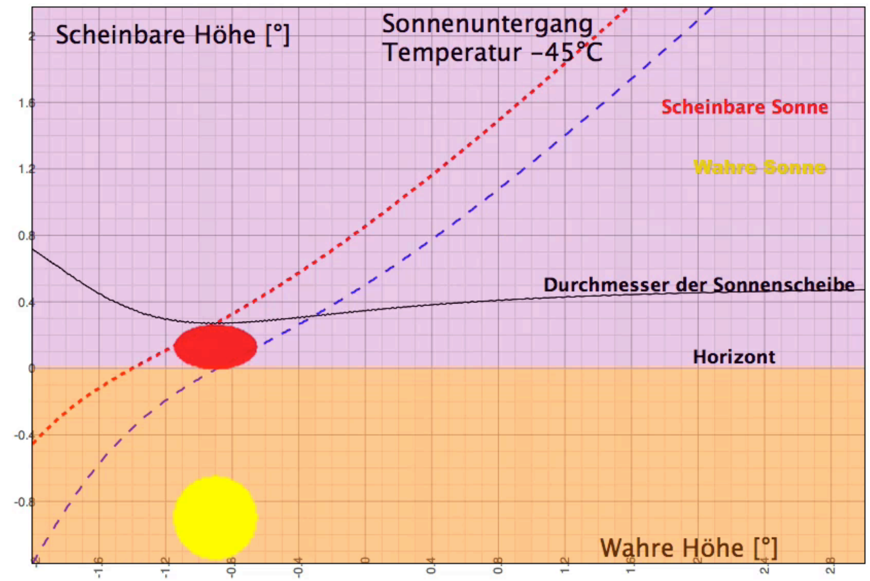
with: I_1 non-damped intensity
I_2 damped intensity

Compare to the optical depth:

$$\tau = \ln\left(\frac{I_1}{I_2}\right)$$

Refraction

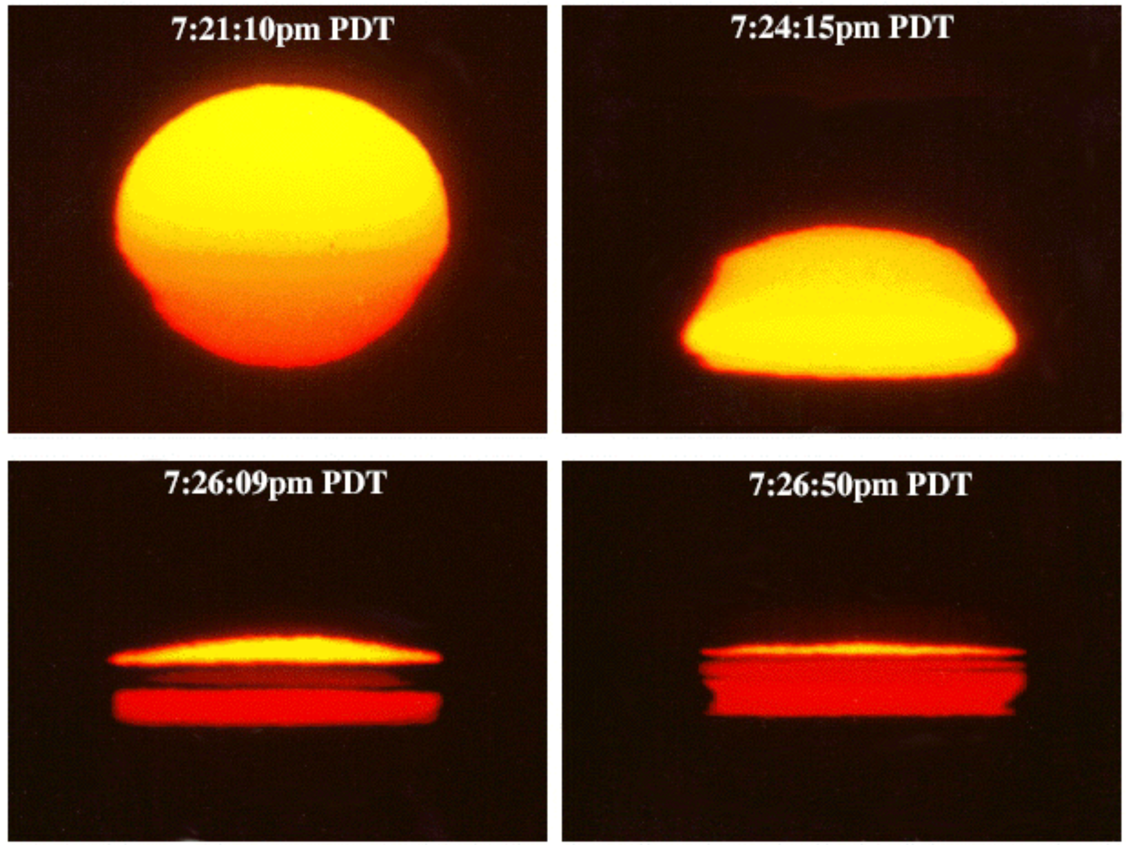
- The density of the atmosphere decreases logarithmically, hence the refractivity
- The trajectories of the (refracted) rays are thus curved: The radiation source seems to have a smaller zenith distance than it (really) has.
- The difference grows with the zenith distance and can be up to 1 degree!



Total Refraction and Mirage

Close to the horizon, total reflection at **inversion layers** may occur:

h=40’ h=2‘



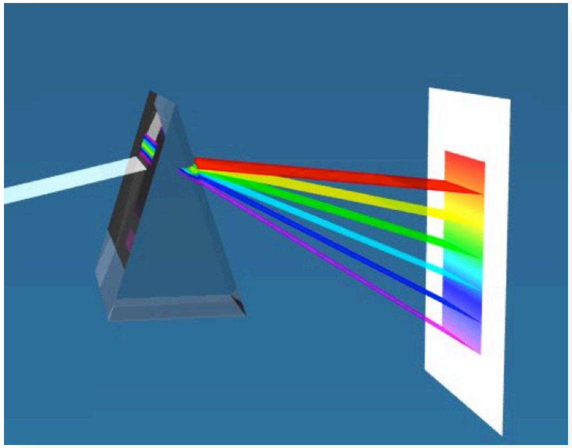
h=-20’ h=-29’

Dr. Andrew T. Young
from Torrey Pines,
north of La Jolla

Fata Morgana

Another phenomena based on total reflection on inversion layers:

Dispersion



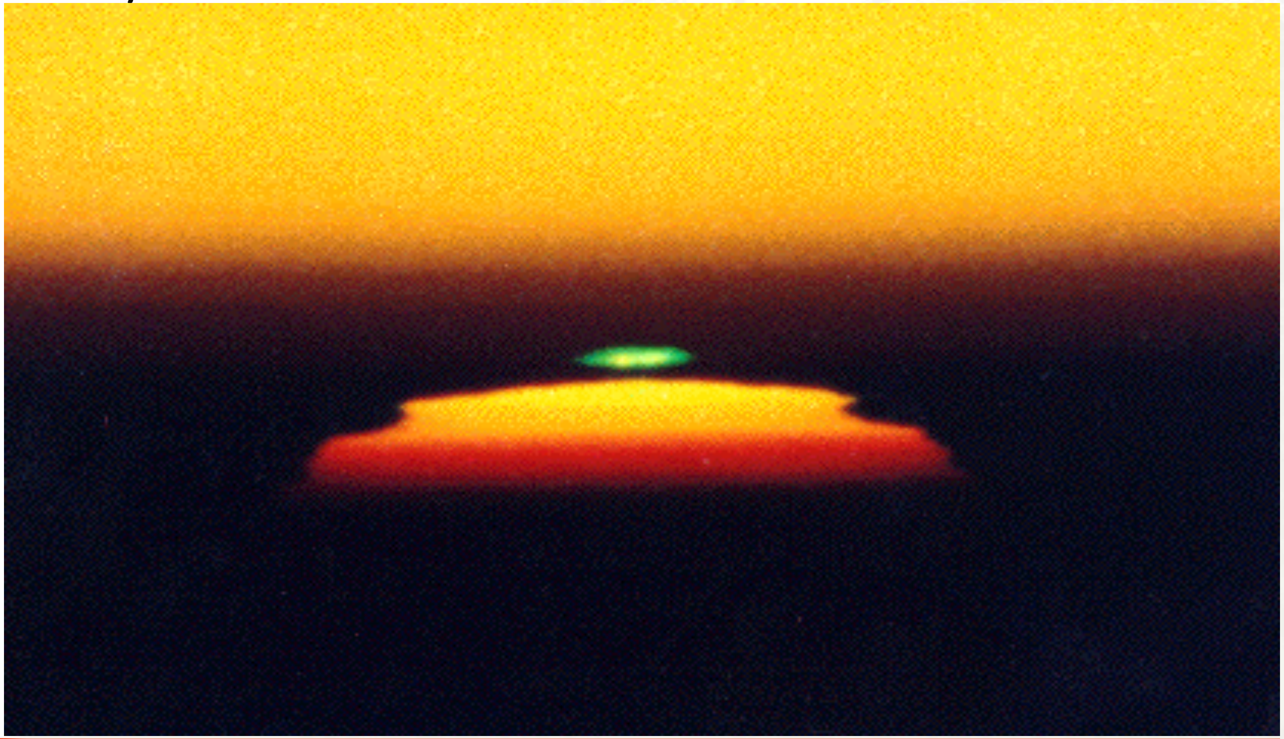
Light of short wavelengths is refracted light of long wavelengths

Rainbow:

Dispersion on water drops

Green Ray

- Due to the dispersion, the light of short wavelengths seems to come from a higher altitude than the light of longer wavelengths
- Under good weather conditions, this can be observed at sunrise and sunset as the green ray:

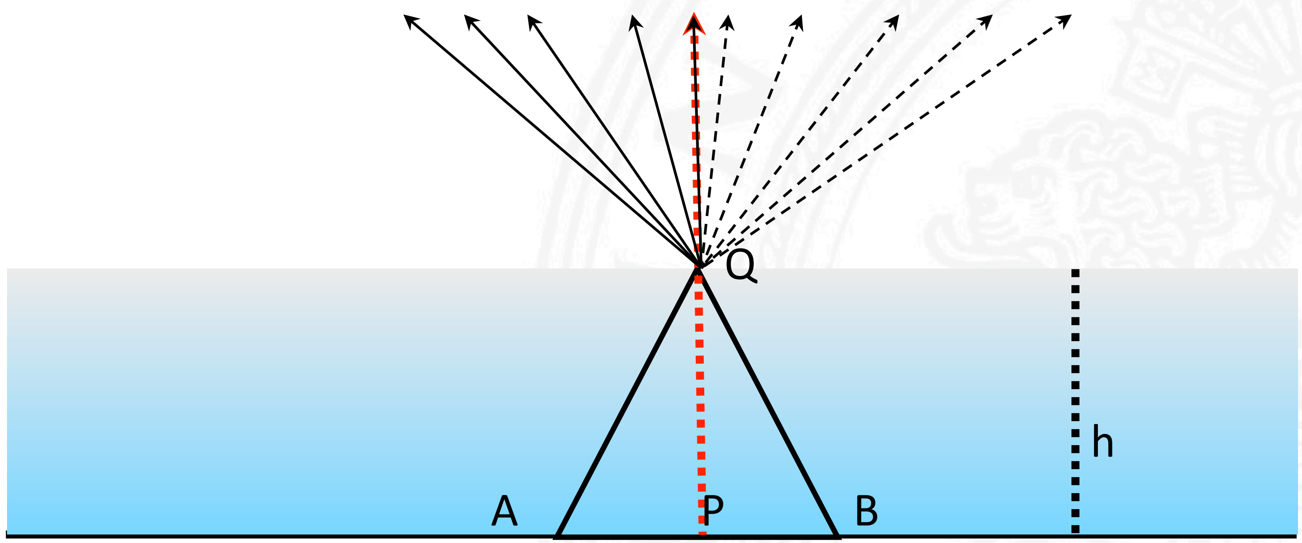


Torrey Pines,
California on Jan. 7, 1996

Turbulences at Stratosphere and Ionosphere

Local changes of density and refractivity due to ascending clusters of warmer gases

- The apparent direction of the radiation may vary randomly.
- Measurement error on ground: 0.02 m
- For current sensor technology: negligible!



Clouds

- About half of the earth surface is permanently covered by clouds
- Unit: 1 octa: 1/8 of the sky is cloud covered
- Clouds restrain Remote Sensing in the visible range of the spectra due to full (opaque) absorbance
- Other spectral ranges may suffer from clouds, too
- For satellites with low revisit rate, it may be problematic to acquire at least one cloud free image!
- Example: Landsat (revisit time 16 days):
  - UK: Estimated only 1 cloud free image per year,
    12.5% cloud coverage: 2 images per year
  - USA: Probability for less than 10% clouds on image: between 5 – 40%

Radiation Trajectories

The **global radiation** B can be composed of two different radiation trajectories:

- **Direct (parallel) sun radiation** I**:**
  The amount of sun radiation, which directly irradiates the terrain.
- **Downward sky radiation** D**:**
  Himmelsstrahlung
  The diffuse light, coming from all directions, scattered inside the atmosphere. Also includes reflection at clouds.

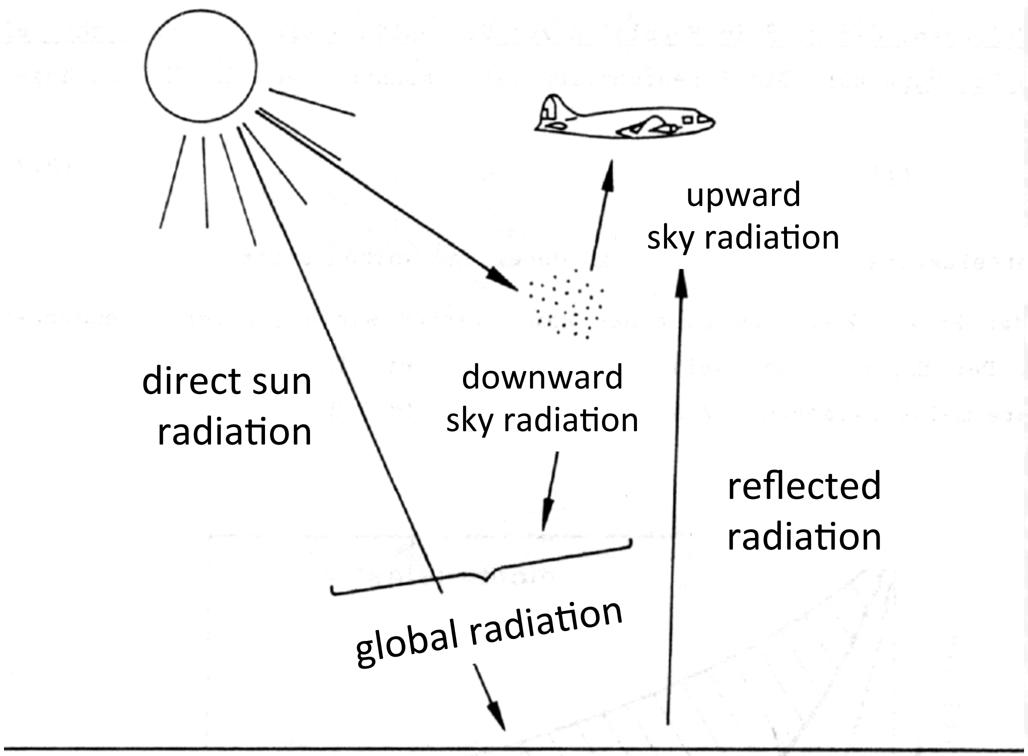
The **upward sky radiation** L (*Luftlicht*) is the part of the diffuse light, with a ascending trajectory.

Global Radiation

- At clear sky and at mean latitudes with a sun in azimuth:
  - Radiation flux of the sun at ground my be up to: $54\ \mathrm{kJ\ m^{-2}\ min^{-1}}$
  - amount of downward sky radiation: 10%
- Extraterrestrial, at $\phi = 50°$ noon at midsummer:
  - $73\ \mathrm{kJ\ m^{-2}\ min^{-1}}$
- On covered sky: $I = 0$, $G = D$, but diffuse radiation can still be quite high:
  - Summer: $33\ \mathrm{kJ\ m^{-2}\ min^{-1}}$
  - Winter: $12.5\ \mathrm{kJ\ m^{-2}\ min^{-1}}$

Radiation at Sensor

- At the sensor, we measure:
  - reflected parts of the global radiation
  - upward sky radiation
- The upward sky radiation overlays the reflected light and thus reduces the contrast!



Directed Reflection and Irradiance

- The correspondence between the directed reflectivity of the terrain surface ρ_r and the Irradiation at the image center E_B is given by:

$$E_B = \frac{d_L^2}{4f^2}\left(E_G \rho_r \tau_A(h, \theta = 0) + \pi L_L(h, \theta = 0, \phi)\right)$$

Terrain reflectivity

Irradiation density of upward sky irrad.

with:

$$E_G = E_{GS} + E_{GH}$$

Upward sky irradiation

Transmissivity of the atmosphere

Direct irradiation

and:

$$E_{GS} = \frac{I_S}{r_S^2}\cos(\theta_S)\left(\tau_A(h = \infty, \theta = 0)\right)^{\frac{1}{\cos(\theta)}}$$

Direct radiation

Compare with Lecture 4, Slide 9!

Example: Irradiation of Terrain (1)

Acquisition altitude 2000 m

Parameter	Ort	Spectral irradiation intensity [$W \cdot m^{-2} \cdot \mu m^{-1}$]	
		λ=0.55µm	λ=0.8µm
Reflection ratio ρ	Acker	15%	25%
	Wasser	5%	0%
	Nadelwald	4%	30%
Direct sun radiation I		$4.43 \cdot 10^{25}\ W \cdot sr^{-1} \cdot \mu m^{-1}$	$2.25 \cdot 10^{25}\ W \cdot sr^{-1} \cdot \mu m^{-1}$
Upward sky radiation E_{GH}	h=2000m, Θ=0	20	10
Downward sky radiation E_{GS}		250	95
Optical Depth τ_A	h=∞	72%	87%
	h=2000	83%	92%

from Kraus Schneider, 1990

Example: Irradiation of Terrain (2)

Spectral Irradiation intensity at the image center (using the equation from slide 37):

	Wavelength	acre			water			conifer forest		
	[µm]	☼	☁	☼/☁	☼	☁	☼/☁	☼	☁	☼/☁
without upward sky radiation	0.55	1.41	0.25	*5.6*	0.47	0.08	*5.9*	0.38	0.07	*5.7*
	0.80	1.53	0.17	*9.0*	0.01	0.0	*6.0*	1.84	0.21	*8.8*
with upward sky radiation	0.55	1.91	0.75	*2.6*	0.97	0.58	*1.7*	0.88	0.57	*1.5*
	0.80	1.78	0.43	*4.2*	0.26	0.25	*1.0*	2.09	0.46	*4.5*

☼=Sun, ☁=Shadow, ☼/☁ Ratio of brightest vs. darkest parts

Note: The contrast ☼/☁ is proportional to the reflection ratio and inverse proportional to the amount of upward sky radiation.

Acquisition Parameters

- The acquisition parameters are assumed to be better, if the monitored objects are imaged at high contrast w.r.t. the background
  - Contrast
  - Horizontal visible range
- Perfect acquisition impossible!
- Atmospheric Correction is based on these parameters!

Contrast of a Scene

- The contrast of a scene is defined as the ratio from which a black object can be distinguished from the background:

$$C = \frac{L_H - L_O}{L_H}$$

with:

L_H Radiance of the upward sky radiation,

L_O Radiance of the object

Horizontal Visible Range v_h

- A black object shows up brighter the higher the upward sky radiation between the object and the observer
- The horizontal visible range v_h is defines as the distance, in which a black object can be distinguished from the sky with a contrast of at least 2%:

$$v_h = \frac{3.91}{ext_\lambda}$$