Lecture, Summer term 2011, LMU München

An Introduction to Global Atmospheric Modelling

Lecture 2: Introduction Physical Climate Models & Fundamentals Atmospheric Radiation Budget

Content:

- **PART 1: Introduction Physical Climate Models**
- **PART 2: Basics of Atmospheric Radiation**
- **PART 3: Numerical formulation in Global Models**
- **PART 4: Daisy World**

Veronika Eyring
DLR - Institut für Physik der Atmosphäre
Oberpfaffenhofen

DLR Institut für Physik der Atmosphäre

Outline

- **Lecture 1 (27 April):** Introduction to Global Atmospheric Modelling
- **Lecture 2 (04 May):** Physical Climate Models & Fundamentals Atmospheric Radiation

Preliminary schedule:

- **Lecture 3 (11 May):** Fundamentals: Chemistry (Aerosols and Gas-phase Chemistry)
- **Lecture 4 (18 May):** Fundamentals: Dynamics of the Atmosphere
- **Lecture 5 (25 May):** Introduction into NCAR Command Language (NCL)
- **Lecture 6 (1 June):** NCL practice (Dr Mattia Righi) **Doppelstunde**
- **Lecture 7 (8 June):** moved to 1 June
- **Lecture 8 (15 June):** Steps in Model Formulation: Example EMAC
- **Lecture 9 (22 June):** Model Evaluation and Uncertainties in Climate Projections
- **Lecture 10 (29 June):** International Climate Modelling Activities: Part 1
- **Lecture 11 (6 July):** EXAM
- **Lecture 12 (13 July):** International Climate Modelling Activities: Part 2
- **Lecture 13 (20 July):** Summary

Final Exam: 6 July 2012, 8:00-10:00

Climate Models

Simulate behavior of climate system, ultimate objective

- Understand key physical, chemical and biological processes that govern climate
- Obtain a clearer picture of past climates by comparison with empirical observation
- Project future climate change

Models simulate climate on a variety of spatial and temporal scales

- Regional climates
- Global-scale climate models – simulate the climate of the entire planet

Processes that must be considered when constructing a climate model

1. **Radiative** - the transfer of radiation through the climate system (e.g. absorption, reflection);
2. **Dynamics** - the horizontal and vertical transfer of energy (e.g. advection, convection, diffusion);
3. **Surface process** - inclusion of processes involving land/ocean/ice, and the effects of albedo, emissivity and surface-atmosphere energy exchanges
4. **Chemical** - for a chemistry-climate model (CCM) or ESM with chemistry
5. **Carbon** – for an ESM (defined as AOGCM + carbon cycle + other components)

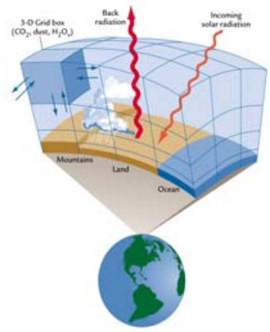
Three-Dimensional Models – AOGCM

3-D representation of Earth's surface and atmosphere

Most sophisticated attempt to simulate the climate system (except ESM)

3-D model based on fundamental laws of physics:

- Conservation of energy
- Conservation of momentum
- Conservation of mass
- Ideal Gas Law



Constructing Climate Models

Basic laws and relationships necessary to model the climate system are expressed as a series of equations which may be

- Empirical derivations based on relationships observed in the real world
- Primitive equations that represent theoretical relationships between variables
- Combination of the two

Equations solved by finite difference methods

- Must consider the model resolution in time and space *i.e.* the time step of the model and the horizontal/vertical scales

All models must simplify complex climate system

- Limited understanding of the climate system
- Computational restraints

Simplification may be achieved by limiting

- Space and time resolution
- Parameterization of the processes that are simulated

Parameterization

Involves inclusion of a process as a simplified function rather than an explicit calculation from first principles

- Sub-grid scale phenomena, like thunderstorms, must be parameterized
  - Not possible to deal with these explicitly
- Other processes are parameterized to reduce computation required

Certain processes omitted from model if their contribution negligible on time scale of interest

- Role of deep ocean circulation while modeling changes over time scales of years to decades
- Models may handle radiative transfers in detail but neglect or parameterize horizontal energy transport
- Models may provide 3-D representation but contain much less detailed radiative transfer information

General Circulation Model (GCM)

Sub-components of the ATMOSPHERIC COMPONENT

Radiation
Aerosols
Clouds
Convection
Precipitation - large-scale and convective
Boundary-layer

Sub-components of the LAND COMPONENT

Vegetation
Soil moisture
Albedo
Energy partitioning
Hydrology

Sub-components of the OCEAN COMPONENT

Absorption of radiation
Salinity variation
Currents
Freezing/thawing near sea ice boundary

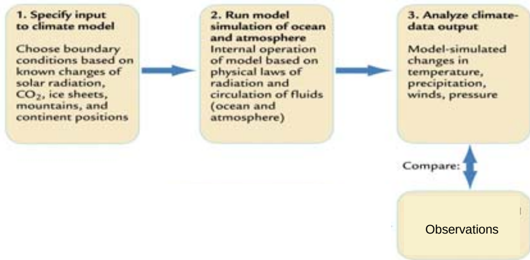
Sub-components of the SEA ICE COMPONENT

Transport of sea ice
Albedo differences
Freezing/thawing near ocean boundary

Data-Model Comparisons

Models constructed to simulate Modern circulation

- Changes based on Earth History inserted in model
- Climate output compared with observations



Climate Physics: Lecture on Radiation Processes

PART 2
Basics of Atmospheric Radiation

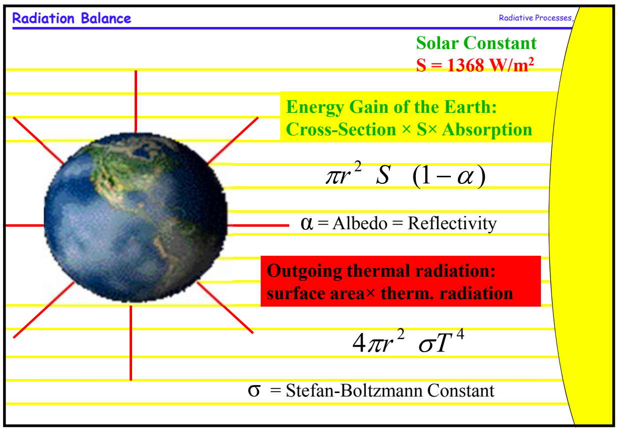
Overview

1 Radiation Laws (Emission)
- Planck's Law
- Wien's Displacement Law
- Stefan Boltzmann Law

2 Absorption, Scattering, Reflection

3 Radiative Transfer Equation

Radiation Balance



Solar Constant
S = 1368 W/m²

Energy Gain of the Earth:
Cross-Section × S× Absorption

$$\pi r^2 \; S \; (1-\alpha)$$

α = Albedo = Reflectivity

Outgoing thermal radiation:
surface area× therm. radiation

$$4\pi r^2 \; \sigma T^4$$

σ = Stefan-Boltzmann Constant

Natural Greenhouse effect

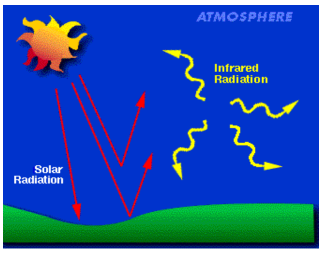
$$\pi r^2 (1-\alpha) S = 4\pi r^2 \sigma T^4$$

Radiation Temperature:

$$T = \sqrt[4]{S(1-\alpha)/(4\sigma)}$$

Earth:
S = 1368 Wm^{-2}
α_e = 0.3
Therefore:

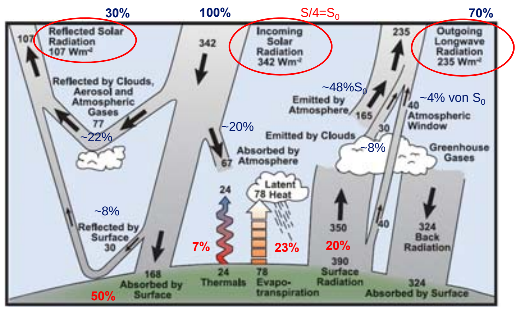
T_e = 255K = -18°C



The natural greenhouse effect causes the mean temperature of the Earth's surface to be about 33°C warmer than it would be if natural greenhouse gases were not present.

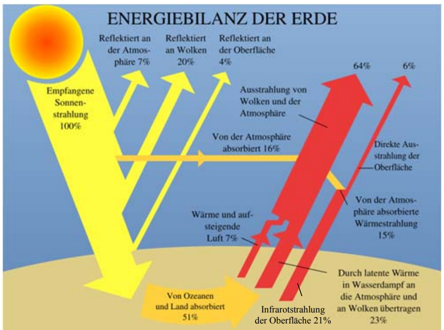
What controls climate? (the simple version)

Estimate of the Earth's annual and global mean energy balance



IPCC AR4

What controls climate? (the simple version)



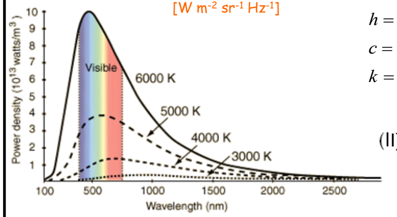
Planck's Law

(I) $$L_\nu^*(T) = \frac{2h\nu^3/c^2}{\exp(h\nu/kT)-1}$$

Spectral Radiance
spektrale spezifische Ausstrahlung

Planck's Law describes the rate of energy output of a blackbody as a function of frequency / wavelength

[W m^{-2} sr^{-1} Hz^{-1}]



$h = 6.625\times10^{-34}\,Js$

$c = 3\times10^{8}\,ms^{-1}$

$k = 1.37\times10^{-23}\,JK^{-1}$

(II) $\nu = c/\lambda;$ $d\nu = -c\lambda^{-2}d\lambda;$ $L_\nu d_\nu = L_\lambda d_\lambda$

(III) $$L_\lambda^*(T) = \frac{2hc^2/\lambda^5}{\exp(hc/\lambda kT)-1}$$

[W m^{-2} sr^{-1} m^{-1}]

Wien's Displacement Law

Frequency of maximum emission:

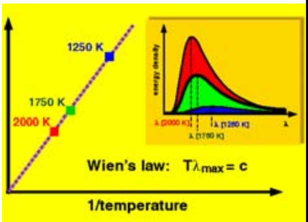
- If the wavelength of maximum emission of the spectral distribution of the black body is plotted over 1/T, one obtains a straight line.
- Wavelength of peak radiation emitted by an object is inversely related to temperature, so
- the hotter an object is the shorter the wavelength at which it will emit most of its radiation

$$\nu_{max}(T) = 0.941\frac{3k}{h}T; \quad \lambda_{max} = \frac{a}{T}$$

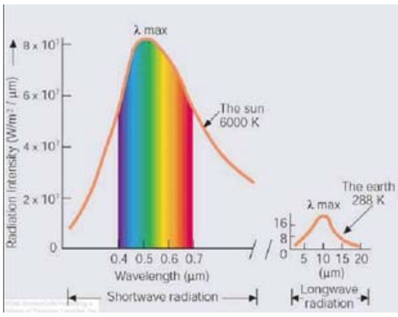
$a = 2.898\cdot10^{-3}\,mK$

Terrestrial: $\lambda_{max} = 9.66\,\mu m\,(IR)$

Solar: $\lambda_{max} = 0.48\,\mu m\,(VIS, green)$



Spectrum of the Sun compared with that of the Earth



The hot sun radiates at shorter wavelengths that carry more energy.

Energy absorbed by the cooler earth is then re-radiated at longer wavelengths, as predicted by Wien's displacement law.

Stefan Boltzmann Law

Stefan-Boltzmann law:

- Total irradiance of a black body from integration of Planck's Law over the entire wavelength domain
- The flux density emitted by a blackbody is proportional to the fourth power of the absolute temperature

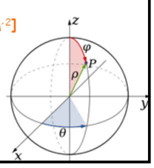
$$L(T) = \int_0^\infty L_\lambda d\lambda = \frac{2k^4\pi^4}{15c^2h^3}T^4$$

- Integrating over all solid angles using polar coordinates (dΩ=sinθ dθ dφ):

$$M(T) = \int_\Omega L(T)\cos\theta d\Omega = \int_0^{2\pi} d\varphi \int_0^{\pi/2} d\theta \sin\theta\cos\theta L(T) = \sigma T^4$$

[W m^{-2}]

$$\sigma = \frac{2k^4\pi^5}{15c^2h^3} = 5.67\cdot10^{-8}\,Wm^{-2}K^{-4}$$



Interaction between radiation and materia

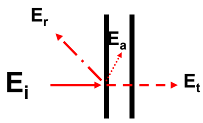
Definitions:

absorptivity α = amount of absorbed radiant energy (E_a) / total amount of incident energy (E_i)

emissivity ε = α (Kirchhoff's Law)

reflectivity ρ = reflected radiant energy (E_r) / total amount of incident energy (E_i)

transmissivity τ = transmitted radiant energy (E_t) / total amount of incident energy (E_i)



$E_i = E_a + E_r + E_t$

Conservation of Energy:
ρ + α + τ = 1

Examples:		
	black body	ε = 1
	white body	ρ = 1
	grey body	0 < ε < 1

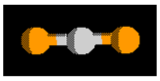
Absorption of Radiation by Molecules

Rotational modes:

- atoms of a molecule rotate around axis
- important for absorption of longwave radiation
- usually requires low energies

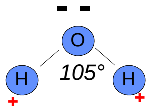
Vibrational modes:

- atoms of a molecule "wobble"
- important for absorption of longwave radiation
- occurs at higher energies than rotational bands



Rotational and vibrational modes

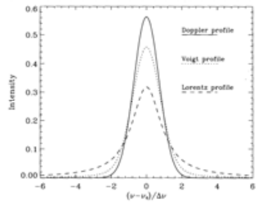
- require a permanent dipole moment (uneven distribution of electrons) to produce oscillating electric dipole moment and therefore affection transmission of electromagnetic radiation
- mainly found in molecules with three atoms, e.g. H_2O, CO_2



Line Broadening

Instead of discrete lines, transitions are observed in a whole wavelength region:

- **Natural broadening:** due to **Heisenberg's uncertainty principle**, limited importance
- **Pressure broadening (Lorentz):** due to collision of molecules or atoms, most important
- **Doppler broadening:** due to Doppler effect (**thermal motion**), important at lower pressure

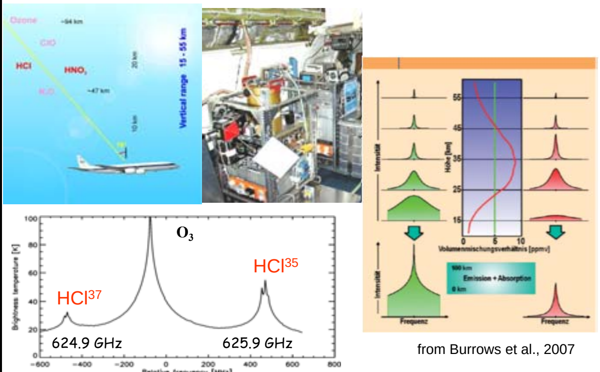


Width of discrete spectral lines in the atmosphere (rotational modes)

Doppler:	$\gamma \approx 1$ MHz	**Lorentz:**	1000 hPa	≈ 2000 MHz
			100 hPa	≈ 200 MHz
			10 hPa	≈ 20 MHz
			1 hPa	≈ 2 MHz

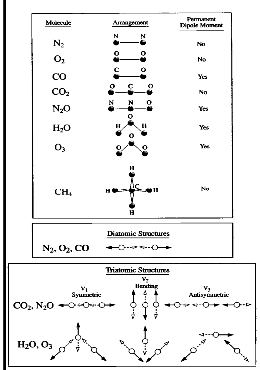
Measurement of Vertical Profiles

Airborne Submillimeter Radiometer (ASUR), measures between 600 - 700 GHz



from Burrows et al., 2007

Molecular Absorbers/Emitters in the IR



- Molecules of gas in the atmosphere interact with photons of electromagnetic radiation
- Different kinds of molecular transitions can absorb/emit very different wavelengths of radiation
- Some molecules are able to interact much more with photons than others

CO_2:

- NIR + vib-rot absorption band near 15 μm (very important for climate as it occurs near the peak of the terrestrial spectrum)

HO_2:

- NIR (between 1 and 4 μm)
- vib-rot band near 6.3 μm
- densely spaced band of pure rotational lines which strongly absorb terrestrial emission at wavelength in excess of 12 μm

O_3:

in the middle of the water vapor window at 9.6 μm

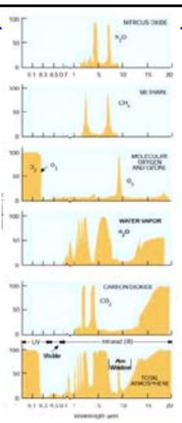
Atmospheric Absorption

Solar radiation passes rather freely through Earth's atmosphere

Earth's re-emitted longwave energy either fits through a narrow "window" or is absorbed by greenhouse gases and re-radiated toward earth

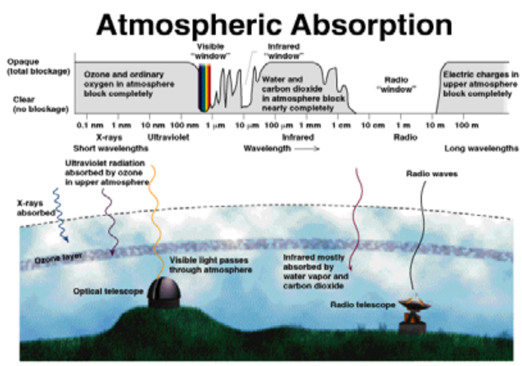
Major LW absorbers:

Water vapor
CO2
O3
Clouds Clouds

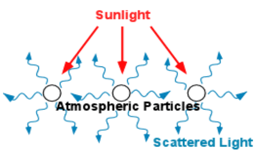


Atmospheric Absorption

Electromagnetic Spectrum

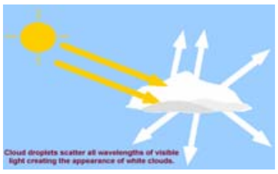


What is scattering ?



Mie scattering:
The size of the scattering particles (d) is comparable to the radiation wavelength (λ). The power of the scattered wave is very sensitive to the ratio d/λ.

Rayleigh scattering:
The scattering particles are much smaller than the radiation wavelength. Shorter waves are scattered more severly (*selective scattering*). The power of the scattered wave is $\sim\lambda^{-4}$

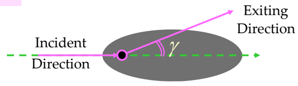


The energy is dispersed in all directions as if the scattering particles act as a new source of radiation.

Scattering Coefficient and Phase Function

$$\frac{dL_S}{ds} = -\alpha_s L$$

α_s : scattering coefficient



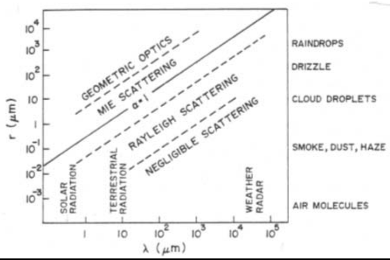
α_s	**Scattering coefficient**
$P(\gamma)$	**Phase function** ➢ probability of light getting scattered in a single direction ➢ Phase function integrates to 1
$\frac{\sigma_s}{4\pi} P(\gamma)$	**Light Scattered in any direction** (σ_s: scattering cross section units m^2)
σ_s	The **scattering cross section** represents the amount of incident energy which is removed from the original direction due to a single scattering event.

Mie parameter

Depending on the ratio of the size of the scattering particle (r) to the wavelength (λ) of the light:

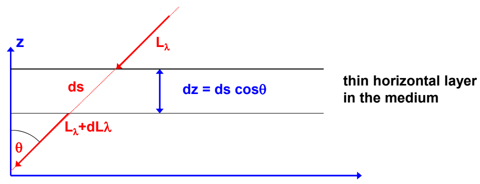
Mie parameter $\alpha = 2\pi r / \lambda$,

Different regimes of atmospheric scattering can be distinguished:



Radiative Transfer Equation (1)

Fundamental equation describing the propagation of electromagnetic radiation in a scattering and absorbing medium



$L_\lambda = dL/d\lambda$ = spectral radiance **[$Wsr^{-1}m^{-3}$]**

forward

Radiative Transfer Equation

Components of Radiative Transfer Equations

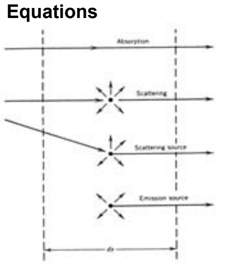


Figure 6-11. The different elements that contribute to the radiation transfer equation.

Wave attenuation due to absorption

$$(1) \quad \frac{dL_A}{ds} = -\alpha_a L$$

Loss of energy in the observer's direction due to scattering

$$(2) \quad \frac{dL_{S1}}{ds} = -\alpha_s L$$

Energy is added in the observer's direction as a result of scattering of waves incident from other directions

$$(3) \quad \frac{dL_{S2}}{ds} = -\alpha_s \bar{J}(\theta, \phi)$$

Energy is added due to thermal emission

$$(4) \quad \frac{dL_e}{ds} = -\alpha_a B(T)$$

Radiative Transfer Equation: Loss Terms

Radiative Transfer Equation – Loss Terms

$[-\alpha_a(\lambda) - \alpha_s(\lambda)]\, L(\lambda, \theta, \phi, s)$
1 2

1 **sum of the absorption coefficients of all the gases and particles in the medium** [m^{-1}]
2 **Scattering coefficient** [m^{-1}]

extinction coefficient: $\epsilon^*(\lambda) = \alpha_a(\lambda) + \alpha_s(\lambda)$

$\epsilon^* = \epsilon\,\rho = \rho\,(\kappa + \sigma)$ [m^{-1}]

ρ	**density of the medium**	[$kg\ m^{-3}$]
ε	**mass extinction coefficient**	[m^2kg^{-1}]
κ	**mass absorption coefficient**	[m^2]
σ	**mass scattering coefficient**	[m^2]

If mixture of gases:

$$\kappa_\lambda = \sum_i \kappa_\lambda^{(i)} \frac{\rho^{(i)}}{\rho}$$

$$\sigma_\lambda = \sum_i \sigma_\lambda^{(i)} \frac{\rho^{(i)}}{\rho}$$

Radiative Transfer Equation – Source Terms

$\alpha_s(\lambda)\,\tilde{J}(\lambda,\theta,\phi) + \alpha_a(\lambda)\,B(\lambda,T)$

1 (Scattering) 2 (Emission)

1 **Scattering** $\tilde{J}(\theta,\phi) = \frac{1}{4\pi}\int_0^{4\pi} \mathbf{L}(\theta',\phi')\,\mathbf{P}(\theta',\phi',\theta,\phi)\,d\Omega'$

scattering phase function

The scattering phase function describes the angular distribution of scattered field

(θ, φ) ⇒ incident radiation

(θ', φ') ⇒ scattered radiation

2 Emission: Planck's Law

$$\frac{dL_\lambda}{ds} = [-\alpha_{a,\lambda} - \alpha_{s,\lambda}]L_\lambda + \alpha_{s,\lambda}\tilde{J}_\lambda + \alpha_{a,\lambda}B_\lambda$$

Losses

Sources

The solution at every point in the medium depends on the interactions between the radiation and the medium at every other point.

$$\frac{dL_\lambda}{ds} = -\tilde{\varepsilon}_\lambda L_\lambda + J_\lambda(z,\theta,\varphi)$$

where J is the sum of both source terms.

The solution of the radiative transfer equation is generally rather complex and can usually only solved numerically. Simplifications exist only for very specific applications and spectral regimes.

$$\alpha_{\nu_{ij}} = N\,\sigma(\nu_{ij},T)\,S(\nu_{ij},p,T)$$

N: Number Density [Particles cm^{-3}]

$VMR = \frac{p_{partial}}{p_{ges}}$ $N = \frac{n}{V} = \frac{p_{partial}}{kT} = \frac{VMR \cdot p_{partial}}{kT}$

σ : Absorption Cross Section [cm^2 Particles^{-1}]

JPL Catalogue $\sigma(vi_j, T)$

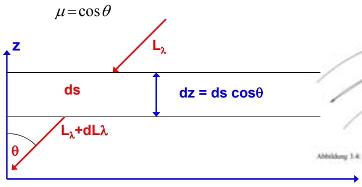
S : Line Broadening

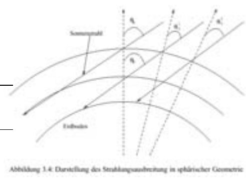
- **Natural Broadening** $\Delta E \Delta t \approx \frac{h}{2\pi}$ $\Delta t \approx \frac{h}{2\pi h \Delta \nu} \approx \frac{1}{2\pi \Delta \nu}$
- **Doppler Broadening** Second major source of line broadening. Molecules are in motion when they absorb. This causes a change in the frequency of the incoming radiation as seen in the molecules frame of reference
- **Pressure Broadening** Line width depends on the number of collisions per second,i.e. on the number density of the molecules (Pressure) and the relative speed of the molecules (the square root of the temperature)

Moleküle besitzen für elektromagnetische Wellen einen Absorptionskoeffizent. Der Absorptionskoeffizienten ist wellenlängenabhängig.
Die absorbierten Lichtwellen erzeugen in den Gasmolekülen:
Rotation, Vibration => Wärme
angeregte Zustände => Lichtwellen anderer Wellenlänge
Der Absorptionskoeffizient wird in 1/m angegeben; d.h. welcher Anteil der Intensität pro Meter Gas (bei bestimmten Druck) absorbiert wird.

Plan-parallel (SZA < 75°)

$\mu = \cos\theta$





- **Formal Solution of the RTE**
- **Vertical integration**
  - i.e. accounting for the variations in **temperature**, **pressure** and density of radiatevely **active absorbers** and **scatterers**
  - gives a directional quantity called **monochromatic radiance**
- **Integration over solar zenith angle**
  - gives a **monochromatic irradiance**
- **Spectral Integration**
  - i.e. an integration over the relevant part of electromagnetic spectrum (LW,SW)
  - gives a **total flux**
- **Differentiation of the Vertical Component**
  - to get the radiative heating/cooling rate

Mote and O'Neill

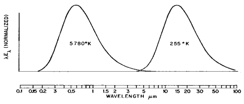
The object of any parameterisation of atmospheric radiation for use in an atmospheric circulation models is to provide a simple, accurate and fast method

- **Calculation must supply:**
  - Total radiative flux at the surface (to calculate the surface energy balance)
  - Vertical and horizontal radiative flux divergence to calculate radiative heating and cooling rates of an atmospheric volume
- **Parmertization should include**
  - absorption and scattering by absorbing gases, clouds and haze particles
- **Trade off between accuracy and speed**
  - Level of approximation and level of speed determine the interactions between radiation and dynamics
  - Radiation affects dynamics
  - Dynamics repsond to the total heating fields (sum of latent and radiative heating and sensible heating components)

Stephens

Simplifications of radiative computations in global models

- E.g., Plane-parallel assumption
- Gas constituents are homogeneously mixed within each cell
- Separtation between Shortwave (SW) and Longwave (LW) radiation schemes owing on the obvious difference between a black-body at the Sun's temperature (T~5800K) and that of the atmosphere (T~255K seen at TOA)
  - SW: most of ist energy below 4μ
  - LW: most of ist energy above 4μ



- Computation of radiation less frequently called than dynamics
  - typical radiation time step: 2 hours
  - typical dynamical time step: 30 minutes

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Heating Rates

- SW and LW separated on the basis of wavelength
  - computation in LW and SW in spectral bands
  - LW: scattering is neglected
- Computing of radiative heating and cooling rates
  - usually done by calculations of upward and downward fluxes through unit horizontal areas, taking into account the vertical distributions of temperature, water vapour and other radiatively active gases (e.g. CO_2, O_3).
  - Total Flux: $F_{rad} = F_{LW} + F_{SW}$
  - Radiative heating rates **Q** are calculated from the difference of the total flux F_{rad} at the lower and upper boundary of a cell, the amount of air m and the specific heat c_p of moist air

$$Qrad = -(F_{rad}^{lower} - F_{rad}^{upper}) / (m \cdot c_p)$$

$$c_p(t) = (1 - q_v(t)) \cdot c_{pd} + q_v(t) \cdot c_{pv})$$

$$m = (p_{lower} - p_{upper}) / g$$

ECHAM5 / Washington and Parkinson

Photolysis Rates

See also Lecture on Chemistry

First order reaction

A + hν -> C + D

Time Dependence for A (J is the photolysis frequency [s^{-1}]):

$$\frac{d[A]}{dt} = -J[A].$$

Example:

$$O_2 + h\nu\ (\lambda < 242\text{nm}) \longrightarrow 2\,O.$$

Photolysis Reaction Rates

Species A undergoes photodissociation.

Reaction *i*: A + hν → products

$$\left.\frac{d[A]}{dt}\right)_{reac\ i} = -[A] \cdot \int \sigma_X(\lambda)\, \phi_i(\lambda, T, \ldots)\, E(\lambda, \ldots)\, d\lambda$$

- A absorption cross section → $\sigma_X(\lambda)$
- Reaction *i* quantum yield → $\phi_i(\lambda, T, \ldots)$
- Wavelength → $d\lambda$
- Spectral actinic flux → $E(\lambda, \ldots)$
- Action spectrum: $\sigma_X(\lambda)\, \phi_i(\lambda, T, \ldots)$
- A concentration rate of change due to photolysis reaction *i*: the left-hand side
- Reaction rate coefficient j: the integral

$$J = \int_\lambda \sigma(\lambda)\, \varphi(\lambda) \int_\omega I(\lambda, \theta, \phi)\, d\omega\, d\lambda$$

Quantum Yield

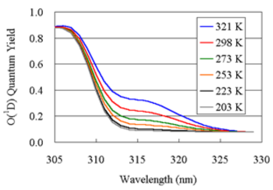
Example for Quantum Yield:

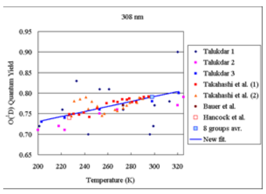
$\Sigma\phi_i = 1$, where summation is over all possible pathways

$\phi_A = \Sigma\phi_i$, where summation is over all pathways that yield A

Photolysis of O_3 + hν (λ < 325 nm) -> $O(^1D)$ + O_2

The quantum yield for O(1D) production in the photolysis of ozone in the ultraviolet region as a function of wavelength and temperature is a key input for modeling calculations in the atmospheric chemistry.

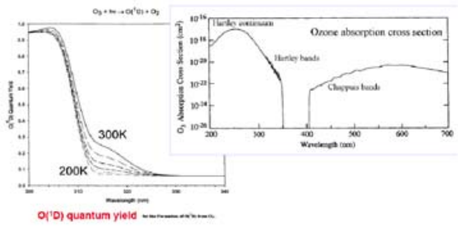




http://www.sparc.sunysb.edu/html/QY_O1D/

Ozone photolysis

Ozone photolysis



$O_3 + h\nu \rightarrow O(^3P) + O_2, \quad \lambda < 800$ nm

$O_3 + h\nu \rightarrow O(^1D) + O_2, \quad \lambda < 320$ nm

Actinic Flux

It is important to distinguish the actinic flux from the spectral irradiance, which refers to energy hitting a flat surface having fixed spatial orientation (J m^{-2} nm^{-1})

- The actinic flux does not refer to any specific orientation because molecules are oriented randomly in the atmosphere.

- This distinction is of practical relevance: the actinic flux (and therefore photolysis) near a brightly reflecting surface (e. g. over snow or above a thick cloud) can be a factor of three higher than that near a non- reflecting surface.

Radiance I_λ Energy flux per solid angle (W m^{-2} sr^{-1} μm^{-1})
Irradiance F_λ: Radiance integrated over all solid angles (W m^{-2} μm^{-1})

Actinic flux

$$F(\lambda) \equiv \int_\omega I(\lambda,\theta,\phi)\,d\omega = \int_\phi \int_\theta I(\lambda,\theta,\phi)\,\sin\theta\, d\theta\, d\phi.$$

(W cm^{-2} nm^{-1} or Photons s^{-1} cm^{-2} nm^{-1})

Actinic Flux

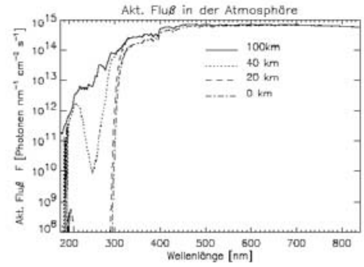
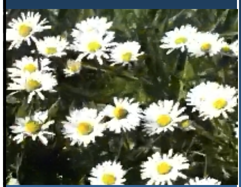


Abbildung 3.2: Aktinischer Fluß innerhalb der Atmosphäre bei 60° Sonnenzenitwinkel, Albedo = 0.3, US-Standard Atmosphäre [*Anderson et al.* (1986)], Sonnenspektrum nach *WMO* (1985).

Global Modelling: Lecture on Daisyworld



PD Dr. habil. Veronika Eyring
DLR - Institut für Physik der Atmosphäre, Oberpfaffenhofen

Overview

Introduction

A new theory of how the world works…

- In 1965, James Lovelock, a atmospheric chemist, was thinking about why life evolved on earth and not on Mars or Venus
- Why has temperature of earth's surface remained in narrow range for last 3.6 billion years when heat of sun has increased by 25%?

Answers:

- Difficult to understood without considering role of life
- We understand that abiotic (non-living) factors (physical, geological and chemical) determine biological outcomes
- New idea is that Biotic (living) factors feedback to control abiotic factors.
- Example of a negative feedback:

Increased Planetary Temperature → Sparser Vegetation, More Desertification → Increased Planetary Albedo → Reduced Temperature

Gaia Theory and Daisyworld

Gaia Theory: Maintenance of Surface Temperatures

- *Gaia: Greek for 'Mother Earth'*: complex entity involving the Earth's biosphere, atmosphere, oceans, and soil.
- According to Lovelock, "Gaia theory predicts that the climate and chemical composition of the Earth are kept in homeostasis for long periods until some internal contradiction or external force causes a jump to a new stable state."
- According to Gaia, life regulates surface temperature because it has remained within 10-20°C for over 3 billion years.
- This is remarkable because the sun's output has increased by 30% or 40%.

Daisyworld: A simple heuristic mathematical model [Watson and Lovelock (1983)]
To demonstrate the principle of biological homeostasis

- Automatic stabilization a planet's temperature in the face of increased solar luminosity through biological feedbacks
- Biological feedbacks arising out of natural selection alone.

Daisyworld Description

A long time ago, in a galaxy far, far away...

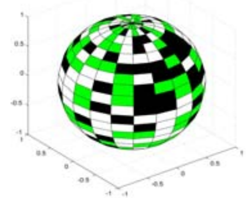
Daisyworld is an imaginary planet

- of same size, rotation, distance from the Sun as Earth
- Sun of the same mass and luminosity as our sun,
- Cloudless, no greenhouse gases, more land than ocean area.
- Fertile, well watered soil, plants will grow anywhere if the temperature is right
- The planet is flat, resulting in similar changes in temperature with changing solar luminosity (energy from the sun) and albedo being experienced simultaneously over its surface, and does not experience any seasonality in climate.
- The composition of the planet's biota is similarly lacking in complexity
  - 2 daisy species (light and dark colored flowers)
  - All daisies are capable of reproducing.
  - Assumed Growth Rate:
    - Below 5°C, no daisies grow,
    - Over 40°C, all daisies die,
    - 22.5°C is optimal for growth of all daisies.

Assumptions

- The rate of population change for both species of daisy depends on the death rate and the potential birth rate for that species, and the amount of fertile land available for growth.
- The birth rate for both species of daisy depends on the local temperature near each daisy type.
- The local temperature depends on the difference between the global and local albedo, and on the global temperature. If the local albedo is large then the local temperature is less than the global temperature.
- The global temperature depends on the luminosity of the Sun and the planetary albedo.
- The planetary albedo is the sum of the local albedo components (i.e., the albedo of the black and white daisies and of the bare ground).
  - Albedo of White Daisies is 0.75 , Black 0.25, and bare ground 0.50.
- By natural selection, the percentage of area covered with black or white daisies varies. This varies total albedo, thus affecting global temperature. Automatic positive and negative feedbacks through natural selection act as a thermostat.

Daisyworld

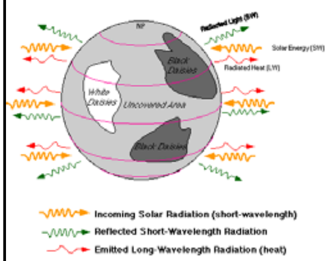


Area	Color	Albedo
F_w	white	α_w=0.75
F_b	black	α_b=0.25
F_g	green (soil)	α_g=0.50

$$F_g + F_w + F_b = 1$$ **total area**

$$\alpha_g F_g + \alpha_w F_w + \alpha_b F_b = \alpha$$ **global albedo**

Schematic Model of Daisyworld



- Black and white daisies
- The temperature is related to how much solar energy is received and how much energy is reflected
- Amount of energy reflected depends on albedo
- The albedo in turn depends on the coverage by white and black daisies.
- The heat radiated to space is a function of how much energy is absorbed.

Daisyworld: Population Dynamics

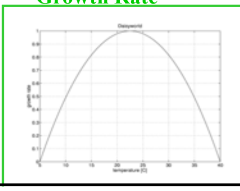
Change of White Area $$(I)\quad \frac{dF_w}{dt} = -\gamma F_w + \beta(\theta_w) F_w F_g$$

Change of Black Area $$(II)\quad \frac{dF_b}{dt} = -\gamma F_b + \beta(\theta_b) F_b F_g$$

Ad hoc Death Rate per unit area (constant)

Growth Rate

$$\beta(\theta) = 1 - 0.003265\,(22.5 - \theta)^2$$



Daisyworld: Energy Balance

$$\sigma T_e^4 = S(1-\alpha)$$ S=solar constant — **effective temperature of Daisyworld**

$$\sigma T_i^4 + N_i = S(1-\alpha_i)$$ i = w,b,g — **(III) local balance**

Heat transport (from warm to cold, i.e. from black via green to white)

$$N_i = q(\alpha_i - \alpha)$$

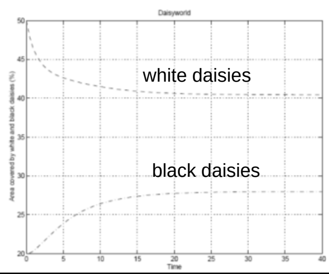
-S < q < 0; q=-S: max heat transport
q=0: zero heat tranpsort

$$\sigma \sum_i F_i T_i^4 = S(1-\alpha) = \sigma T_e^4$$ **(IV) global balance**

$$\alpha_g F_g + \alpha_w F_w + \alpha_b F_b = \alpha$$ **global albedo**

Daisyworld: Numerical Solution

- Now we have all ingredients to solve the system (I), (II), (III) and (IV).
- It is not possible to solve it analytically. Instead, the solution of this rather simple system has to be calculated numerically.
- Solution to the time dependent problem, indicating the evolution towards an equilibrium state



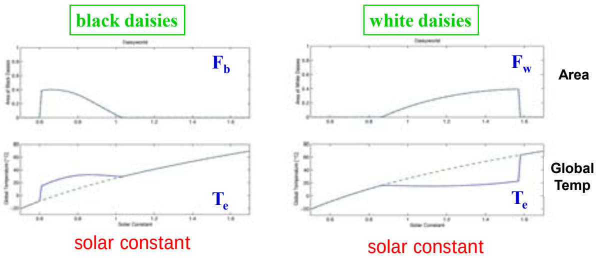
Area in percent

γ = 0.3

S =917 Wm[-2]

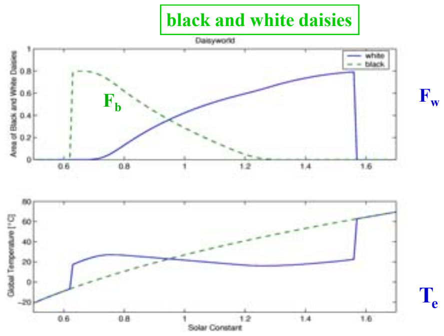
q = -0.9 S

Daisyworld: Energy Balance coupled to Population Dynamics



=> Without a feedback the global temperature is strictly increasing with solar luminosity

Daisyworld: Energy Balance coupled to Population Dynamics



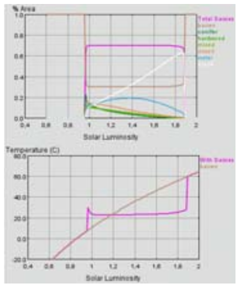
However, the simple feedback in Daisyworld keeps the temperature stable even with increasing S (thermostat). The temperature my even drop as S increases.

Daisyworld: Conclusions

- The **simple feedback** in Daisyworld keeps the temperature stable even if the solar luminosity varies (thermostat). The temperature may even drop as S increases.
- **The black daisies** increase the absorption (and therefore temperature) for a (normalised) solar constant between **0.6 and 1.0 (cold sun),** and the **white daisies** reduce very effectively the temperature for values of the (normalised) solar constant between **1.0 and 1.6 (hot sun).**
- **Below values of 0.6 and above values of 1.6** daisies do not grow.
- **This thermostat-effect** makes Daisyworld a habitable planet for the two species, for a wide range of flux densities of incoming solar radiation.
- In contrast to the positive ice-albedo feedback, the feedback in Daisyworld regulates the temperature. **This negative feedback stabilizes the system.** Positive feedbacks usually have a destabilizing effect.
- In some respects this model is typical for the climate and Earth system. Strongly nonlinear components (here the dynamics of the daisy population) are coupled to positive and negative feedbacks which affect the radiation balance of the planet.
- Because of the nonlinearity of the system there are multiple solutions (equilibria).

Erweiterungen des Daisyworldmodells: Thermostat

- Spätere Erweiterungen des Daisyworldmodells schlossen sogenannte Kaninchen, Füchse und andere Arten mit ein, welche Absorptionsraten zwischen denen schwarzen und weißen Daisys haben.
- Eines der mehr überraschenden Ergebnisse dieser Simulationen war, dass **je größer die Anzahl der Arten war, desto größer die selbstregulierenden Kräfte** des gesamten Planeten. Dies unterstützte die Ansicht, dass Biodiversität wertvoll ist, und löste die moderne Biodiversitätsdebatte aus.
- Daisyworld zog auch eine Reihe Kritik auf sich. Es weist kaum Ähnlichkeit mit der Erde auf; das System benötigt eine **Ad-hoc-Todesrate (γ)** um im Gleichgewicht zu bleiben und das Modell verwischt die Unterschiede zwischen Phänomenen auf der Ebene der Arten und jener der Individuen. Jedoch zeigt Daisyworld unbestreitbar, dass biologisch reguliertes Gleichgewicht keine teleologische Erklärung benötigt.



Further Reading and Exercise

Animation for Daisyworld:

.\gaia1_Dateien\gaia1.htm

Further Reading:

- Kirchner, 1989
- Lovelock, 1989
- Watson and Lovelock, 1983
- Dennis Hartmann, Global Physical Climatology, 1994

Further Exercise:

- http://cs.clark.edu/~mac/physlets/DaisyWorld/Daisy.htm?l1=0.75&aw=0.2&ab=0.2
- http://www.gingerbooth.com/courseware/daisy.html

End of Lecture 5