Chapter 4

Climate Change

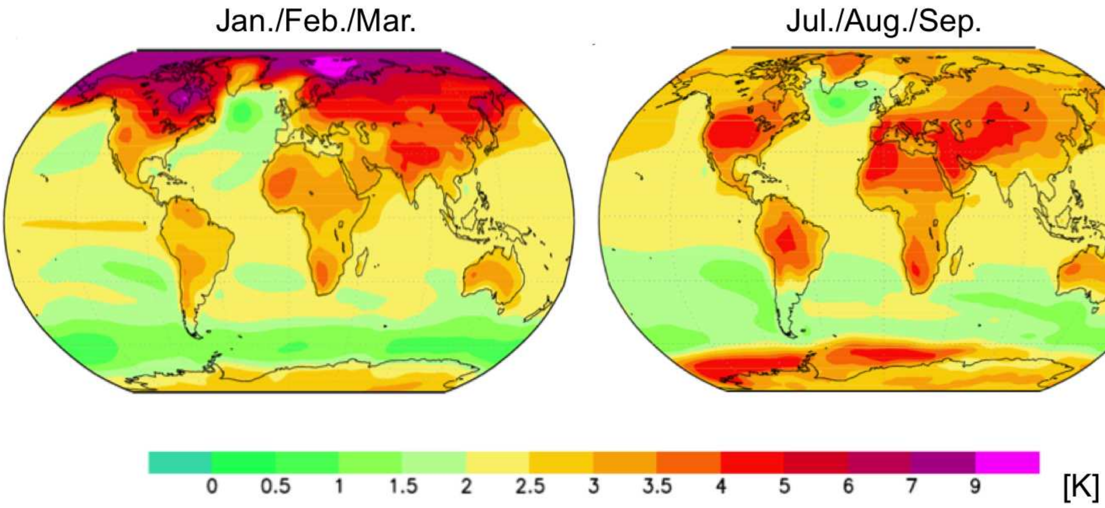

Figure 4.1: T_{surf} response to CO_2 forcing scenario IPCC A1B: Mean of 24 IPCC models (time intervals: [2070-2100] - [1970-2000])

The anthropogenic climate change is arguably the most important aspect of climate dynamics today. It it therefore the central aspect of this course. Before we go into any details it is helpful to have look at the main feature of the anthropogenic climate change: The surface temperature, T_{surf} warming over the 21th century, see Fig. 4.1. Some main features of the global T_{surf} response to CO_2 forcing:

- Global mean warming by the end of the 21st century is about 2.6K.
- There are strong regional differences, although CO_2 concentration is the same everywhere.
- Land warms more than the ocean.
- The Arctic has the strongest warming.
- The Northern Hemisphere warms more than the Southern.
- The winter (cold) season warms more than the summer (warm) season.

The main aim of the next two sections is to understand what physical processes are involved and how they cause these structures. We then discuss the IPCC-type of climate models and their predictions. To understand the significance of the anthropogenic climate change we have to put it into the context of natural current and past climate variability. Finally we will have a look at the discussion of anthropogenic climate change in the media.

4.1 A Globally Resolved Energy Balance (GREB) model

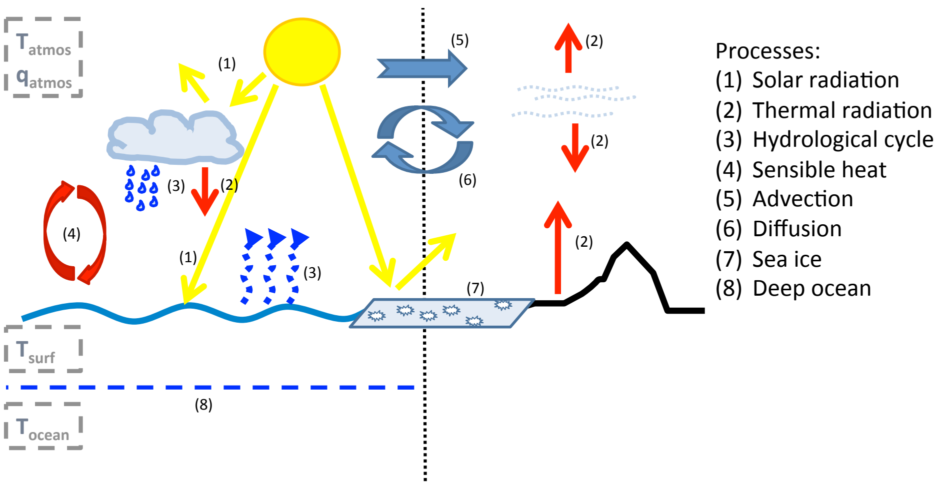

Figure 4.2: Sketch GREB model process:

In this section we want to develop a Globally Resolved Energy Balance (GREB) model to understand the processes involved in the climate change response on the regional scale. The aim of this model is to develop the simplest possible representation of the physical climate system, that can still simulate the main features in the global T_{surf} response to CO_2 forcing shown in Fig. 4.1. So we aim for simplicity not for completeness (see IPCC models for the most sophisticated climate models).

4.1.1 Initial considerations

Some initial considerations:

The energy balance for global mean model is:

$\gamma_{surf}\frac{dT_{surf}}{dt} = F_{solar} + F_{thermal}$

Only radiation exchange to space (no heat or mass transport)

The energy balance for regional model includes more terms, as we can exchange heat with other regions and transfer heat by phase transitions of water (latent heat), see sketch 4.3:

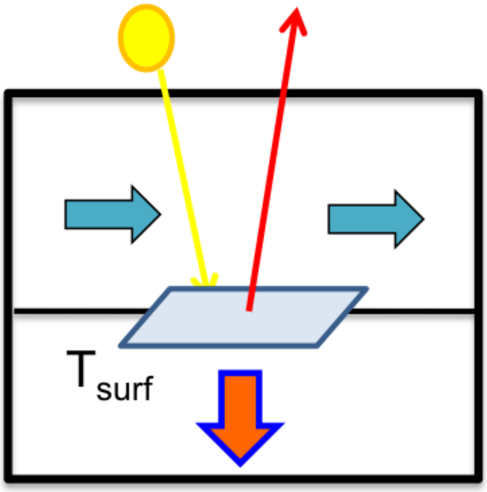

Figure 4.3: Simple sketch of regional heat balances.

$\gamma_{surf}\frac{dT_{surf}}{dt} = F_{solar} + F_{thermal} + F_{sense} + F_{latent} + F_{ocean}$

F_{sense} = turbulent heat exchange with atmosphere
F_{latent} = heat exchange by H_2O phase transitions (e.g. evaporation or condensation)
F_{ocean} = turbulent heat exchange with atmosphere

We want to resolve this equation on a global longitudes $\times$ latitude model grid with points every $3.75^o \times 3.75^o$ (96×48 points), see map in Fig. 4.4 for an illustration of the model grid.

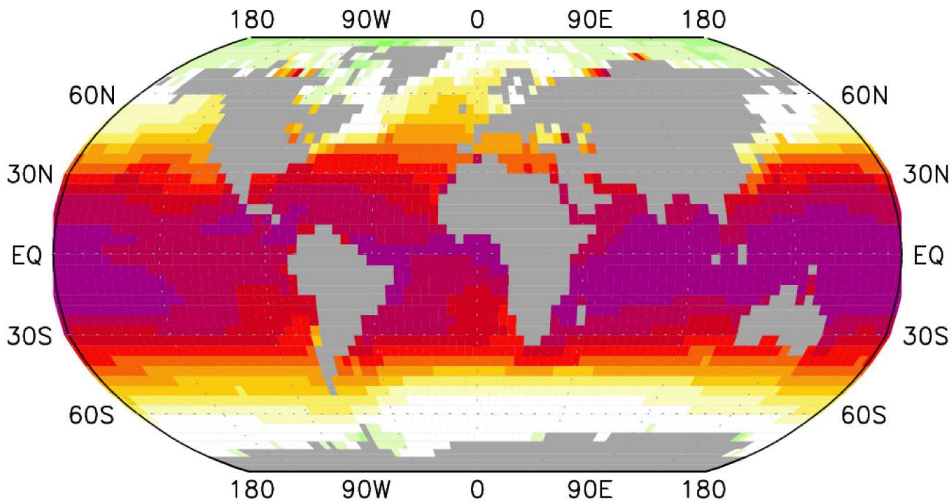

Figure 4.4: Map illustrating the GREB model resolution.

The heat transfer in the climate system is quite complex. It involves heat transport and exchange over many different levels in the atmosphere (see Fig. 4.5). We aim to present the atmosphere with just one layer, neglecting all vertical structure.

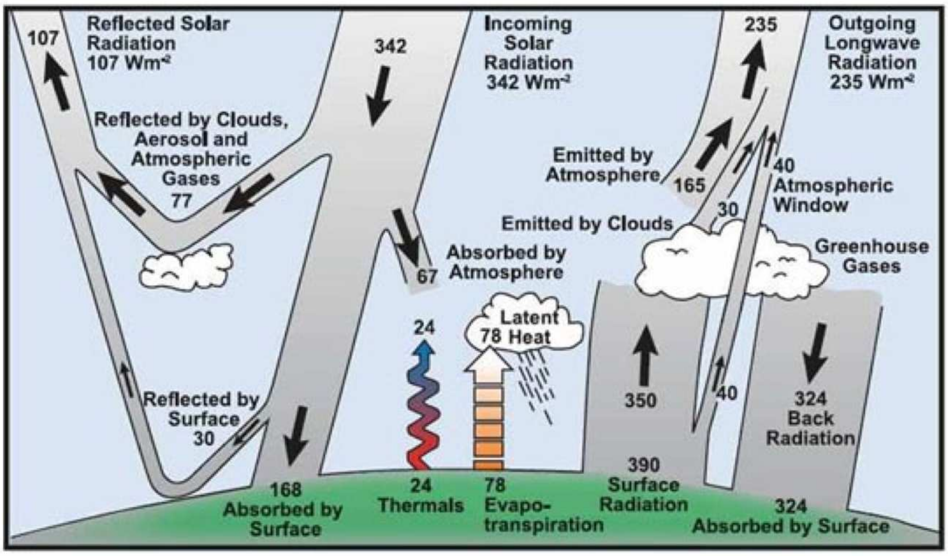

Figure 4.5: A simple sketch of the atmospheric radiation budget. It illustrates the heat exchange in different level in the atmosphere.

4.1.2 Solar radiation (F_{solar})

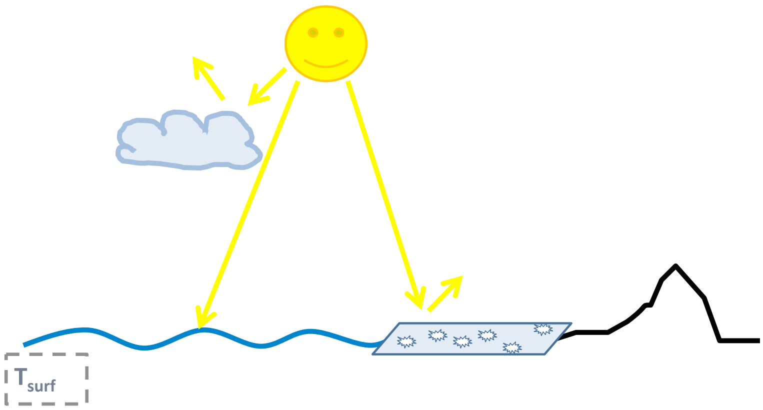

Figure 4.6: Sketch of the GREB solar radiation.

For the global zero order model (e.g. Budyko) the solar radiation was:

$$F_{solar} = \frac{1}{4}(1 - \alpha_p(T_{surf}))S_0 \tag{4.1}$$

For our regional GREB model we have to make it slightly more complex:

$$F_{solar} = (1 - \alpha_{clouds})(1 - \alpha_{surf})S_0 \cdot r(\phi, t_{julian}) \tag{4.2}$$

with $r(\phi, t_{julian})$ 24hrs mean incoming fraction of the solar constant S_0 as function of latitude, ϕ and calendar day of the year, t_{julian}. So in comparison to the zero order model we have included two more aspects in this equation: First, the incoming solar radiation is now given for each day of the year at each latitude and second the reflection of the incoming solar radiation by the albedo is now split into two parts: one part is reflect in the atmosphere by the atmospheric albedo, α_{clouds}, and the remaining incoming radiation is reflected by the surface albedo, α_{surf}.

Incoming Solar Radiation: We make the simplification to assume a 24hrs mean solar radiation, and therefore do not simulate a daily cycle (no nights!). See Fig. 4.7 for the distribution of the 24hrs mean incoming solar radiation as function of latitude, ϕ and calendar day of the year, t_{julian}.

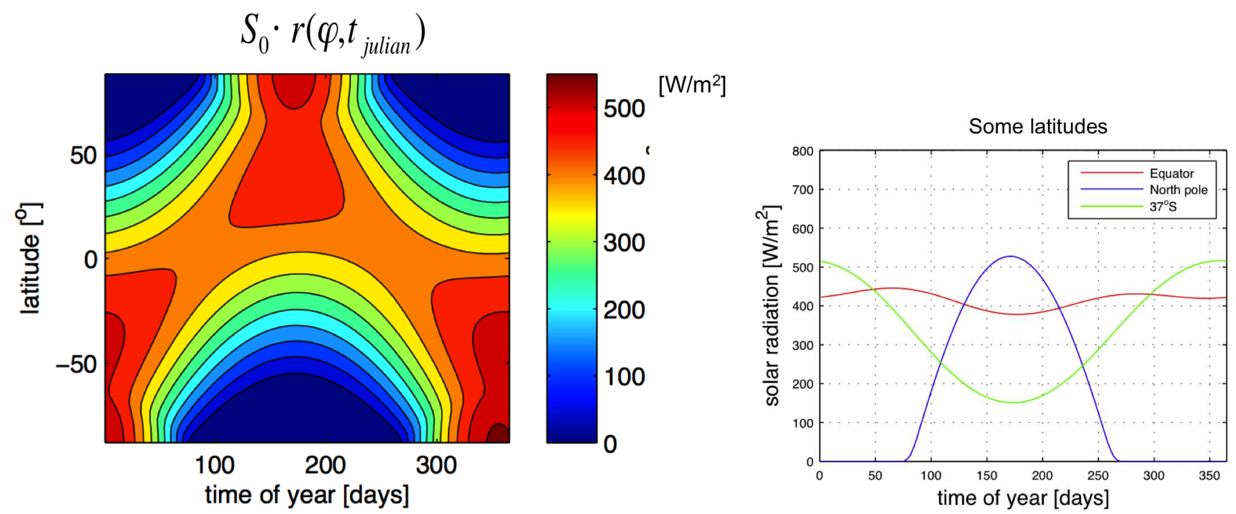

Figure 4.7: 24hrs mean incoming solar radiation as function of latitude, ϕ and calendar day of the year, t_{julian}. The 24hrs mean incoming solar radiation at three different latitudes (the lines are 'cut outs' from the left panel).

Albedo: The atmospheric albedo is

$$\alpha_{clouds} = 0.35 * a_{cloud} \tag{4.3}$$

a_{cloud} = cloud cover
$a_{cloud} = 0 \rightarrow$ no clouds
$a_{cloud} = 1 \rightarrow$ complete cloud cover

So the atmospheric albedo is 0.35 if it is completely cloud covered. Thus it is assumed that clouds reflect about 35% of the radiation. We do not make any differentiation between different type of clouds (e.g. thickness or brightness). We further assume that the cloud cover distribution is a fixed climatology, thus it is not responding to a changing climate. This simplification is not made, because it is realistic (it is probably not realistic). It is made because no simple physical model is known that would tell us how the cloud cover changes for given changes in the climate. In other words: The cloud response is complicated and we first of all do not know how to deal with it. So we assume to first order that there are no changes.

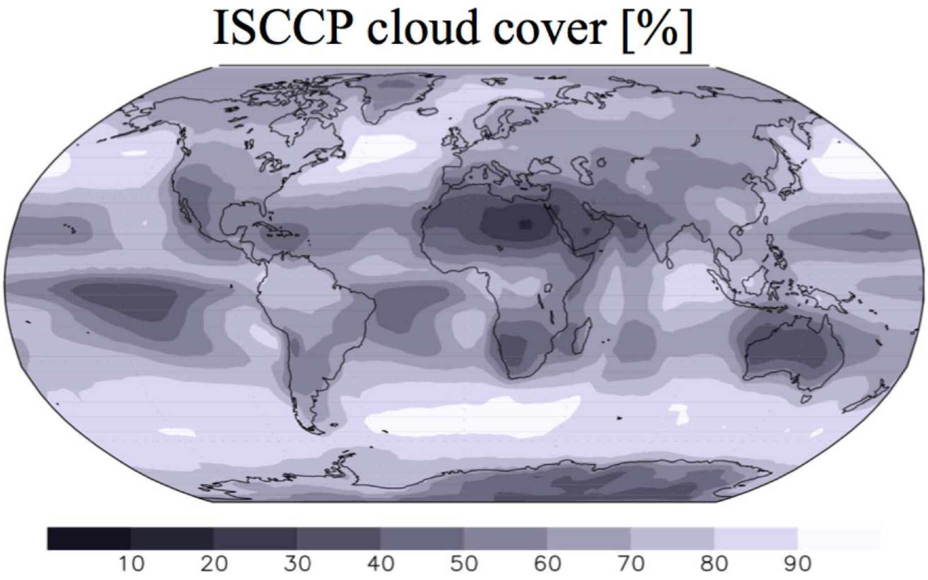

Figure 4.8: ISCCP cloud cover [%]

Surface albedo: The albedo of the earth surface varies with regions, due to different aspects: open ocean waters, snow, ice cover, trees, deserts, etc. See Table 4.1 and Fig. 4.9 for an overview.

Type of surface	**Albedo (%)**
Ocean	2 - 10
Forest	6 18
Cities	14 - 18
Vegetation	7 - 25
Soil	10 - 20
Grassland	16 - 20
Desert (sand)	35 - 45
Ice	20 - 70
Cloud (thin, thick stratus)	30, 60 - 70
Snow (old)	40 - 60
Snow (fresh)	75 - 95

Table 4.1: Albedos for different surfaces on Earth. Note that the albedo of clouds is highly variable and depends on the type and form [from Marshall and Plumb].

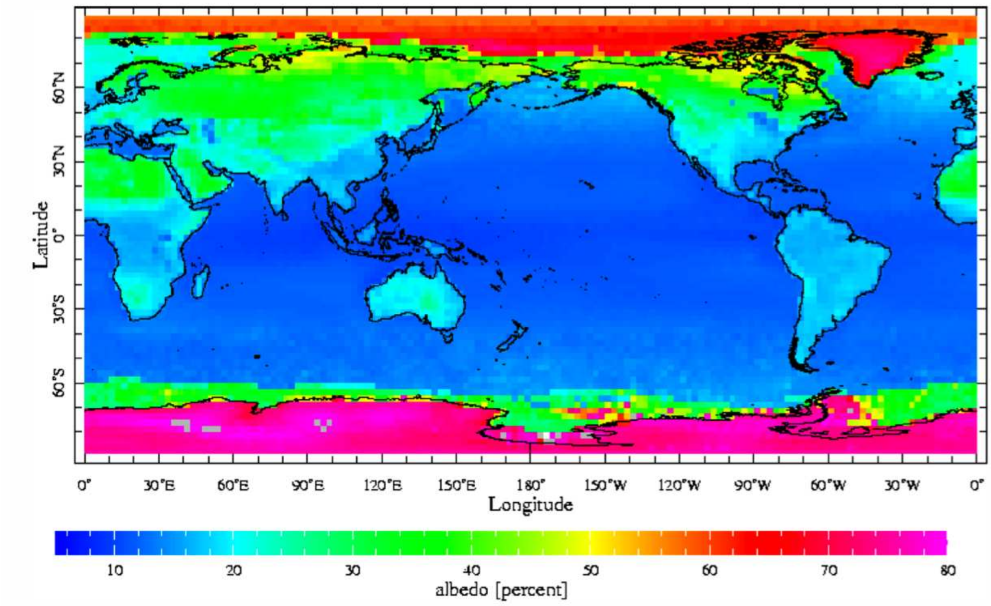

Figure 26: The annual mean surface albedo (%). Source: http://iridl.ldeo.columbia.edu/SOURCES/.NASA/.ERBE

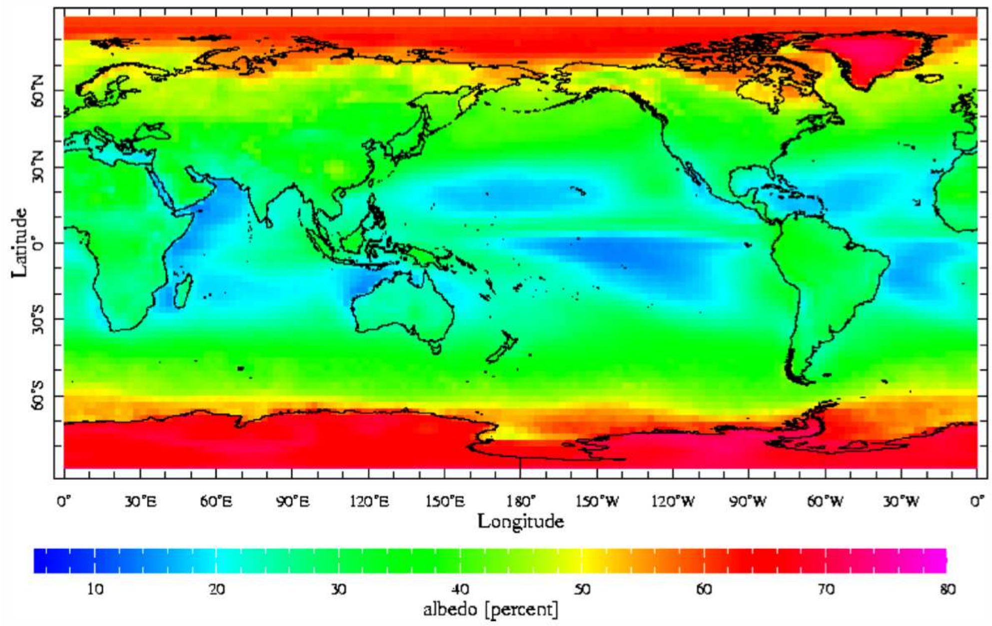

Figure 27: The annual mean surface + atmosphere albedo (%). Source: http://iridl.ldeo.columbia.edu/SOURCES/.NASA/.ERBE

Figure 4.9: Upper: Surface Albedo. Lower: Total albedo of the atmosphere (clouds) and the surface together.

For the surface albedo we assume the same kind of ice-albedo feedback as in the Budyko model (section 2.3). So we assume that the surface albedo is a function of T_{surf}:

$$\alpha_{surf} = \alpha_{surf}(T_{surf}) \tag{4.4}$$

As in the Budyko model we assume a constant albedo for very cold and warm temperatures and an albedo decreasing linearly with T_{surf} in a temperature range slightly below freezing, see Fig.

4.10. Note, that we make a difference between oceans and land temperature ranges. This reflects differences in spread of ice/snow cover response due to differences in freezing points (ocean water at -1.7^oC), heat capacities and other influences, such as cloud cover, topography, land usage, vegetation, dust, etc.

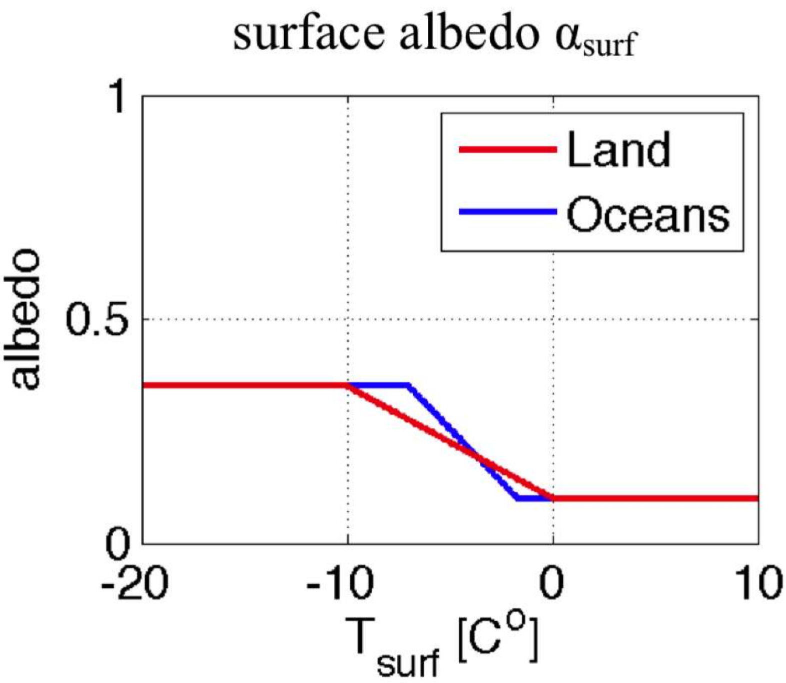

Figure 4.10: Ice-albedo function $\alpha_{surf} = \alpha_{surf}(T_{surf})$

Ice-Albedo feedback: In the temperature range where the albedo is a function of T_{surf} the solar radiation forcing is

$$F_{solar} = (1 - \alpha_{clouds})(1 - \alpha_{surf})S_0 \cdot r(\phi, t_{julian}) \tag{4.5}$$

with

$$\alpha_{surf} = \alpha_0 + \frac{\Delta\alpha}{\Delta T_{surf}} \cdot (T_{surf} - T_0) \tag{4.6}$$

with $\alpha_0 = 0.1$. Over land $\frac{\Delta\alpha}{\Delta T_{surf}} = -0.25/10^oK = -0.025K^1$ and $T_0 = 273.15K$, See fig.4.10. We can now compute the strength of the ice-albedo feedback using the definition of the linear feedback parameter eq. [2.50]:

$$C_{ice-albedo} = \frac{dF_{solar}}{dT_{surf}} = (\alpha_{clouds} - 1) \cdot S_0 \cdot r(\alpha, t_{julian}) \cdot \frac{\Delta\alpha}{\Delta T} \tag{4.7}$$

for $T_{surf} \in [-10, 0]$

We can compute the value for a realistic T_{surf} and cloud cover climatology and the seasonal changing incoming solar radiation, see Fig. 4.11. A few points can be made about this Ice-Albedo feedback:

- It is only active for $T_{surf} \in [-10, 0]$. So regions and seasons where T_{surf} is mostly in this range will have a strong ice-albedo feedback,
- It is strong if solar radiation is strong. No sun light (polar winter) $->$ no ice-albedo feedback. So it stronger at lower latitudes.

- It is weak if cloud cover is large, as it masks the surface.

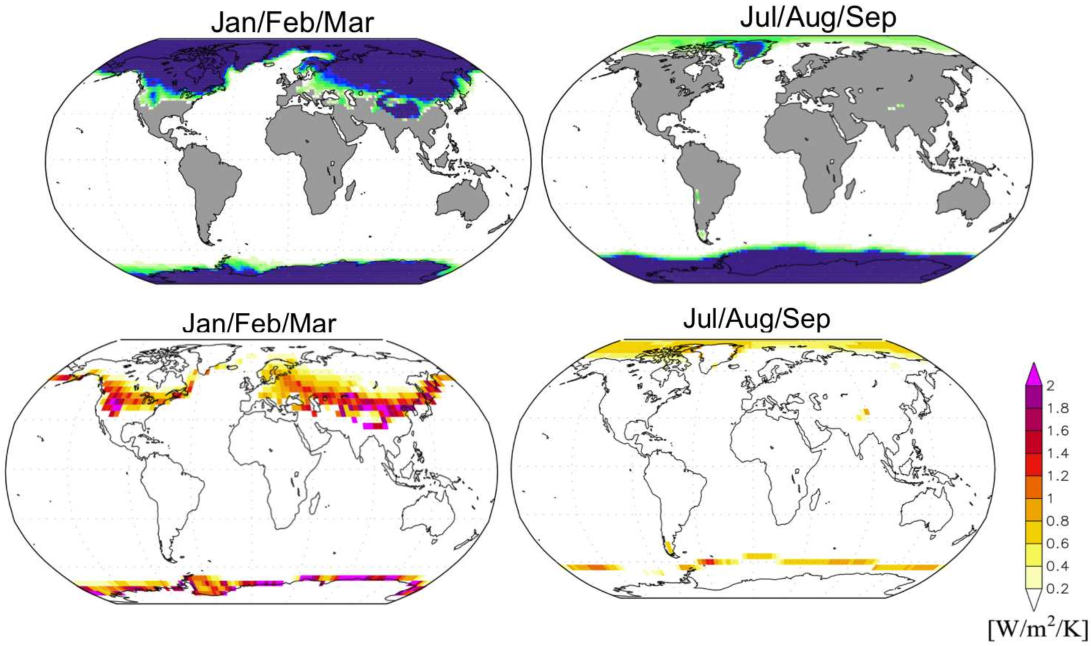

Figure 4.11: Upper: Typical Snow/Ice distribution. Lower: Ice-albedo feedback parameter strength in the GREB model for a given T_{surf}, cloud cover and incoming solar radiation climatology.

4.1.3 Thermal radiation ($F_{thermal}$)

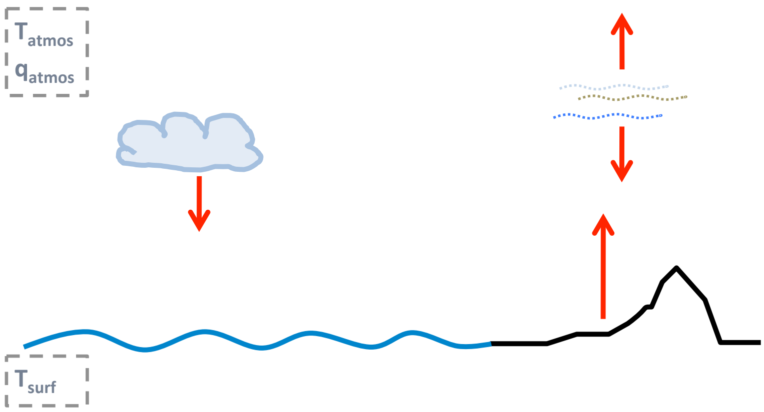

Figure 4.12: Sketch GREB thermal radiation

The thermal radiation of the atmosphere is the **only** way how the CO_2 or the other greenhouse gasses in general influence the climate. It is therefor central in understanding the climate response to anthropogenic forcing. Before we describe the GREB thermal radiation model it is instructive to have a view on a more detailed model of the thermal radiation balance in the atmosphere, see also sketches 4.13 and 4.14:

- Thermal radiation emitted from each layer is a function of emissivity, ϵ, and temperature, T at each layer.
- Thermal radiation absorbed from each layer is a function of the layer's emissivity and that of all other layers ϵ and T.
- Emissivity, ϵ, at each layer is a function of pressure, chemical composition (H_2O, CO_2) and cloud droplet density.
- Emissivity of chemical components is a function of wave length, see Fig. 4.15.
- Overall: .... it's complicated!

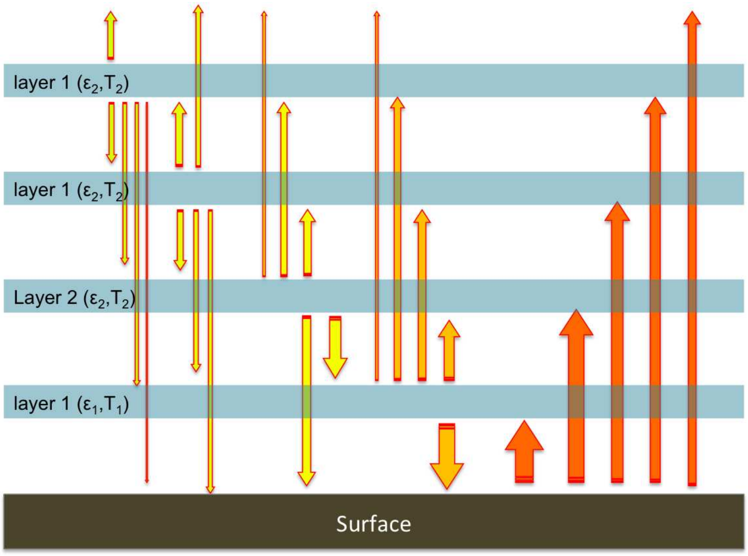

Figure 4.13: A more detailed model of thermal radiation. Each layer absorbs and emits thermal radiation as a function of its emissivity, ϵ_i, and temperature T_i.

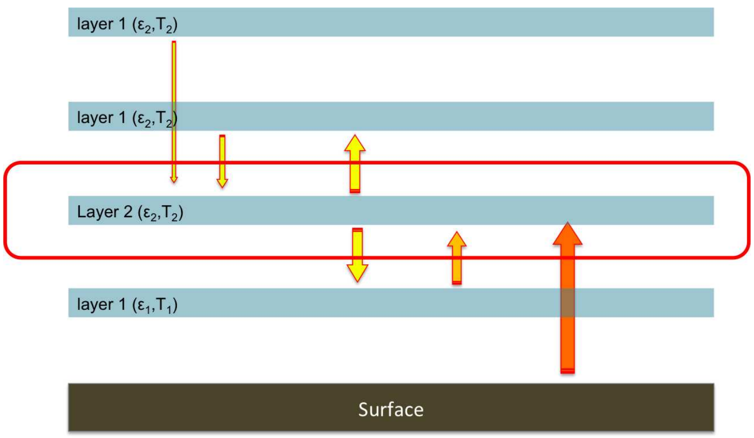

Figure 4.14: Thermal radiation balance for a single layer.

Chemical composition of the atmosphere

Air is a mixture of permanent gases (N_2, O_2) in constant ratio with minor constituents and some non-permanent gases (e.g. water vapor or CO_2).

Chemical species	**Molecular weight (g mol^{-1}**	**Conc (% by vol)**
N_2	28.01	78
O_2	32.00	21
Ar	39.95	0.93
H_2O (vap)	18.02	0.5
CO_2	44.01	385 ppm
Ne	20.18	19 ppm
He	4.00	5.2 ppm
CH_4	16.04	1.7 ppm
Kr	83.8	1.1 ppm
H_2	2.02	500 ppb
O_3	48.00	500 ppb
N_2O	44.01	310 ppb
CO	28.01	120 ppb
NH_3	17.03	100 ppb
NO_2	46.00	1 ppb
CCl_2F_2	120.91	480 ppt
CCl_3F	137.27	280 ppt
SO_2	64.06	200 ppt
H_2S	34.08	200 ppt
AIR	28.97	

Table 4.2: The major atmospheric constituents. The chlorofluorocarbons (CFCs) CCl_2F_2 and CCl_3F are also known as CFC-12 and CFC-11 respectively. (ppm, ppb, ppt) = parts per (million, billion, trillion). Source: Marshall and Plumb (2008).

GREB-model: For the GREB model we use the "greenhouse shield" / "slab atmosphere" model as in the previous Sect. 2.2.2. We simplify the atmosphere to just one layer, see sketch 4.16:

Surface:

$$F_{thermal} = -\sigma T^4_{surf} + \epsilon\sigma T^4_{atmos} \tag{4.8}$$

Atmosphere:

$$F_{thermal} = -2\epsilon\sigma T^4_{atmos} + \epsilon\sigma T^4_{surf} \tag{4.9}$$

GREB emissivity function:

ϵ_{atmos} = F(CO_2, H_2O_{vapour}, cloud cover)

Approach for the emissivity function ϵ_{atmos}:

- Assume saturation effect (a doubling of greenhouse gases does not double the greenhouse effect; see multi-layer greenhouse shield)

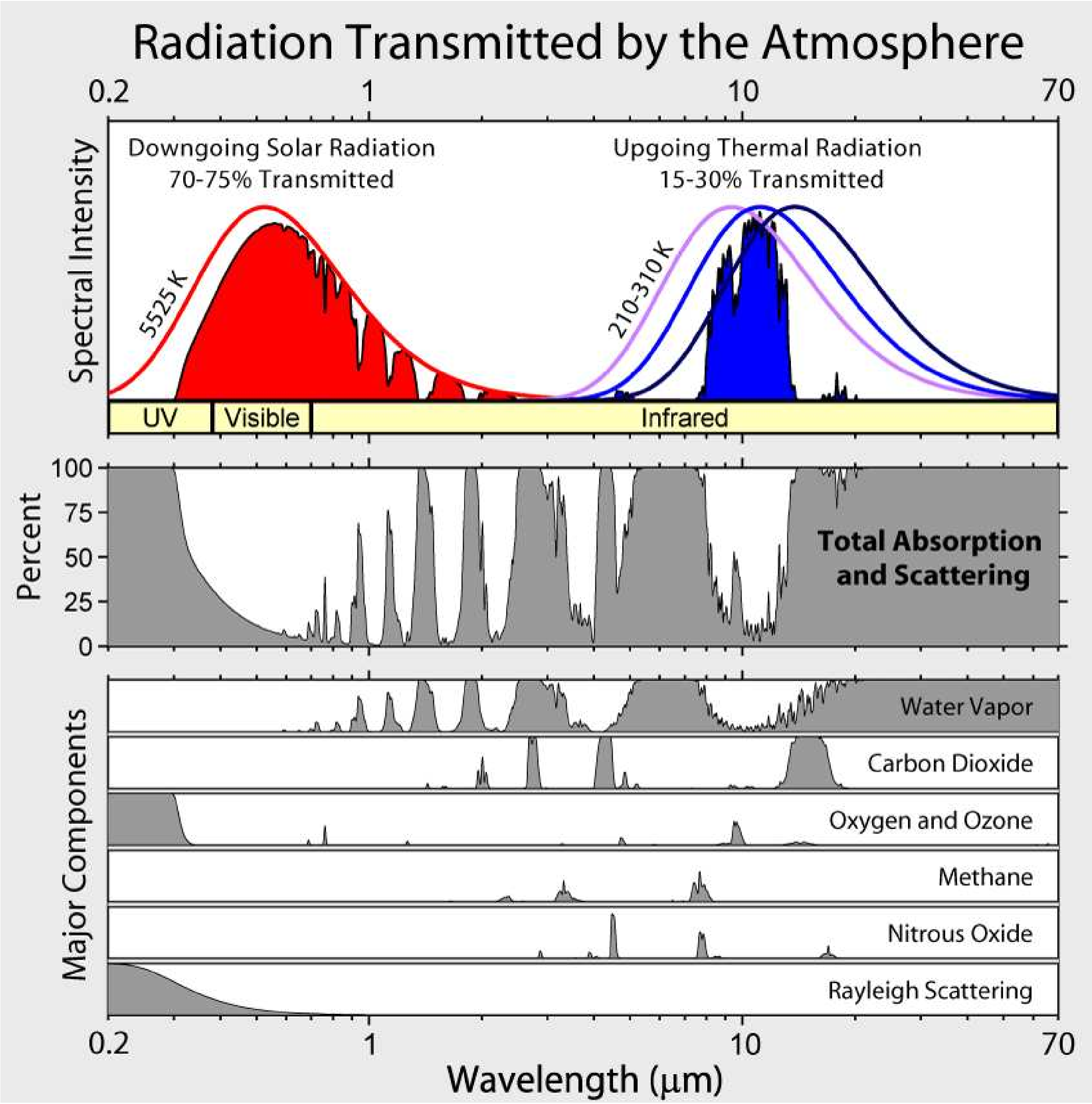

Figure 4.15: Radiation transmitted by the atmosphere.The lower panels: The amount of radiation absorption for the most important chemical components of the atmosphere as function of wave length. Middle: total absorption of the atmosphere as the sum of all components in percentage. Upper: Incoming Solar and out going terrestrial radiation power spectra.

- log-function approximation (to simulate saturation effect - note there is no fundamental physical law behind this log-function)
- 3 spectral bands (to simulate the overlap of absorption bands in H_2O and CO_2)
- Fit function to data (frm more sophisticated radiation models; IPCC-type models)

Saturation of the greenhouse effect: From the Multi layer greenhouse shield model in Sect. 2.2.2:

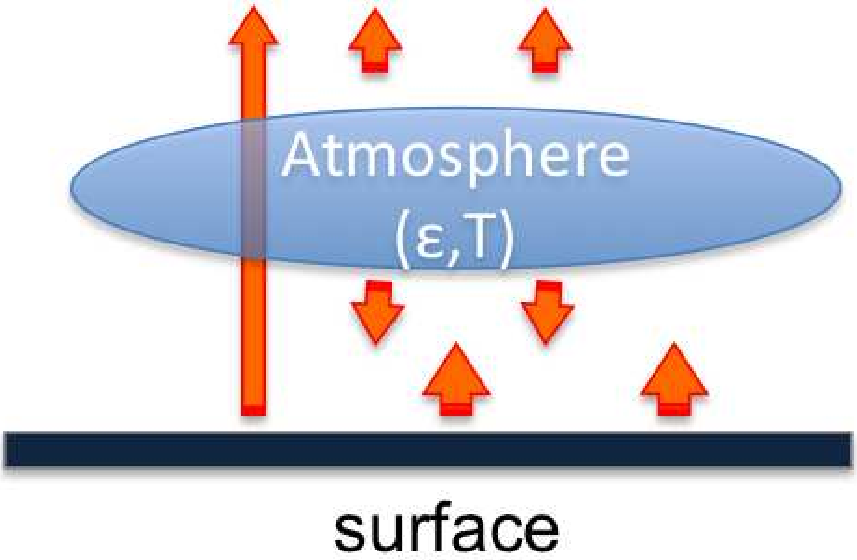

Figure 4.16: Sketch of GREB model thermal radiation: The Greenhouse Shield / Slab Atmosphere model.

$$\rightarrow \sigma T_{surf}^4 = (N+1) \cdot (1-\alpha_p)Q \tag{4.10}$$

$$\rightarrow g = 1 - \frac{1}{N+1} \tag{4.11}$$

From these equation we see that a doubling of concentrations (N) does not double the greenhouse effect. There is a saturation effect.

GREB emissivity function:

ϵ_{atmos} = F(CO_2, H_2O_{vapour}, cloud cover)

$$\varepsilon_{clear-sky} = \underbrace{p_4 \cdot \log\left[p_1 \cdot CO_2^{topo} + p_2 \cdot H_2O_{vapor} + p_3\right]}_{\text{CO}_2\text{ and H}_2\text{O absorb}} \underbrace{+ p_5 \cdot \log\left[p_1 \cdot CO_2^{topo} + p_3\right]}_{\text{only CO}_2\text{ absorbs}} \underbrace{+ p_6 \cdot \log\left[p_2 \cdot H_2O_{vapor} + p_3\right]}_{\text{only H}_2\text{O absorbs}} + p_7$$

Figure 4.17

$\epsilon_{clear-sky}$ = clear sky emissivity (no clouds)

H_2O = total water vapour concentration in air column

$CO_2^{topo} = CO_2 \cdot e^{\frac{z_{topo}}{h_{atmos}}} = CO_2$ concentration in air column

z_{topo} = altitude [m]

h_{atmos} = 8400 m = scaling height

a_{cloud} = cloud cover

Although CO_2 concentration is globally uniform, we need to consider that higher altitudes have less of an atmosphere above.

$\epsilon_{atmos} = \frac{p_8 - a_{cloud}}{p_9}(\epsilon_{clear-sky} - p_{10}) + p_{10}$

$p_4, p_5, p_6 =$ relative importance of each spectral band
$p_1, p_2 =$ greenhouse scaling concentration
$p_3 =$ contribution to greenhouse effect by other gases
$p_7 =$ artificial fitting constant
$p_8, p_9, p_10 =$ cloud scaling parameter
$p_1 > 0$ all parameters are positive

Cloud cover *scales the effective emissivity of ϵ_{atmos} by shifting it up or down and by diluting the effects of the trace gases.*

Characterisitcs of the emissivity function, which we can mostly get from evaluating the radiation spectrum in Fig. 4.15:

1. Global mean emissivity $\epsilon = 0.8$
2. $\Delta\epsilon$ (water vapour $\to 0) \approx 0.5$. the largest part of emissivity is due to water vapour.
3. $\Delta\epsilon(CO_2 \to 0) \approx 0.18$
4. $\Delta\epsilon$(cloud cover$\to 0) \approx 0.16$
5. $\epsilon(CO_2 = H_2O =$cloud cover$= 0) \approx 0$ No greenhouse gases and no clouds $\to$ no emissivity.
6. $\epsilon(H_2O = 70kg/m^2$,cloud cover$= 1) \approx 1$ A very humid and fully cloud covered atmosphere has emissivity of about 1.
7. % of the H_2 absorption is non-overlapping with the CO_2 absorption bands.
8. H_2O and CO_2 have about equal strength in spetral bands where they both absorb.
9. CO_2 absorbs about equally strongly in the two absorption bands.
10. Clear sky sensitivity to greenhouse gases is about twice as strong as for completely cloud covered sky.
11. $\Delta\epsilon(2 \times CO_2) \approx 0.02$ The change in emissivity due to doubling of CO_2, which follows from the IPCC models $3.8W/m^2$ additonal thermal downward radiation.
12. $\Delta\epsilon(\Delta$ water vapour$) \approx 0.02$ It follows from the IPCC models.

- Water vapour has a strong effect on ϵ
- The effect of water vapour is non-linear
- The sensitivity to change in water vapour (slope) is bigger for small amounts of water vapour
- Clouds increase ϵ
- Clouds dilute the effect of greenhouse gases
- The sensitivity to CO_2 is bigger if water vapour is less

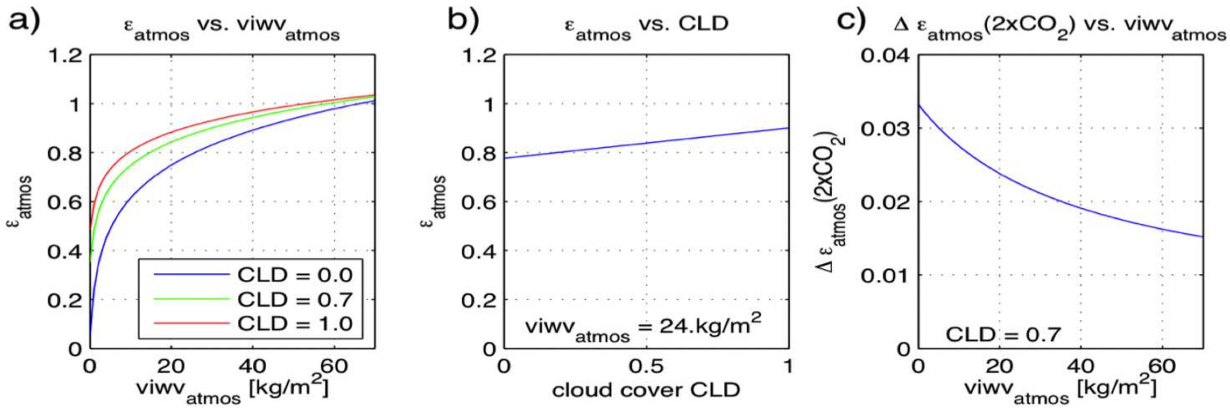

Figure 4.18: GREB emissivity function:

Result of thermal radiation model:

1. We need a prognostic equation for T_{atmos}:
   $\gamma_{atmos}\frac{dT_{atmos}}{dt} = F_{thermal} + \ldots$

2. We need a prognostic equation for q_{atmos}:
   $\frac{dq_{surf}}{dt} = \ldots$

4.1.4 Hydrological cycle

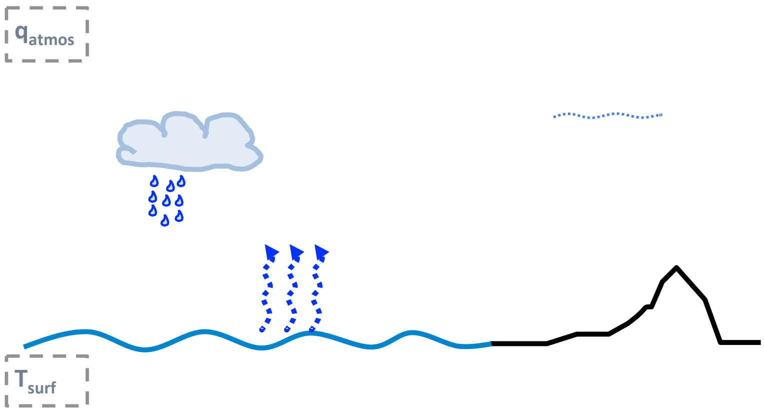

Figure 4.19: Sketch GREB model process: The hydrological cycle; evaporation, precipitation, latent cooling at the surface and heating in the atmosphere and the amount of water vapour in the atmosphere.

In the previous section we have seen that water vapour is the most important greenhouse gas. So need to know the amount of water vapour in the atmosphere. Note, that CO_2 and water vapour have quite different characteristics in the atmosphere:

- Lifetime of CO_2 in atmosphere: 10 yrs - 10,000 yrs (different processes involved).Since it stays in the atmosphere for so long it is globally well mixed and therefore it is globally uniformly distributed; every point on earth on the same pressure level has roughly the same amount of CO_2 in the air above it.
- Lifetime of water vapour in atmosphere: 10 days (rain, weather). So it stays in the atmosphere only very shortly and is therefore not well mixed and has strong regional differences. So we need to simulate it with a prognostic equation:

$$\frac{dq_{surf}}{dt} = \Delta q_{eva} + \Delta Q_{precip} + \ldots \tag{4.12}$$

Δq_{eva} = evaporation of water into the atmosphere
Δq_{precip} = condensation and precipitation of water out of the atmosphere

Complex weather: It rains if water vapour in the atmosphere condenses and the droplets get big enough to fall to ground. Therefore the air must saturate with water vapour. Air typically saturates with water vapour if air is lifted (vertical motion) and therefore adiabatically (by reducing pressure) cooled. Thus many different processes are involved in precipitation and they are controlled by weather fluctuations.
→ Weather fluctuations can not be *'simulated'* in the simple GREB model.

Precipitation:

How to estimate precipitation in GREB?

- Roughly: it rains if there is atmospheric water vapour.
- Atmospheric water vapour stays in the atmosphere for about 10 days

$\Rightarrow \Delta q_{precip} = 0.1 \frac{1}{day} q_{atmos}$

Thus we make the very strong simplification that it rains 10% of the water vapuor in the air column every day.

Evaporation:

- Evaporation at the surface increases the relative humidity, q_{atmos} [%]
- Evaporation or any other phase change of water causes latent heat flux

An empirical bulk formula for the evaporation, Δq_{eva}:

$$\Delta q_{eva} = \frac{1}{r_{H_2O}} \cdot \rho_{air} \cdot C_w \cdot |\vec{u}_*| \cdot v_{soil} \cdot (q_{sat} - q_{atmos}) \tag{4.13}$$

r_{H_2O} = regression parameter
ρ_{air} = density of air at sea level
C_w = empirical transfer coefficient over oceans

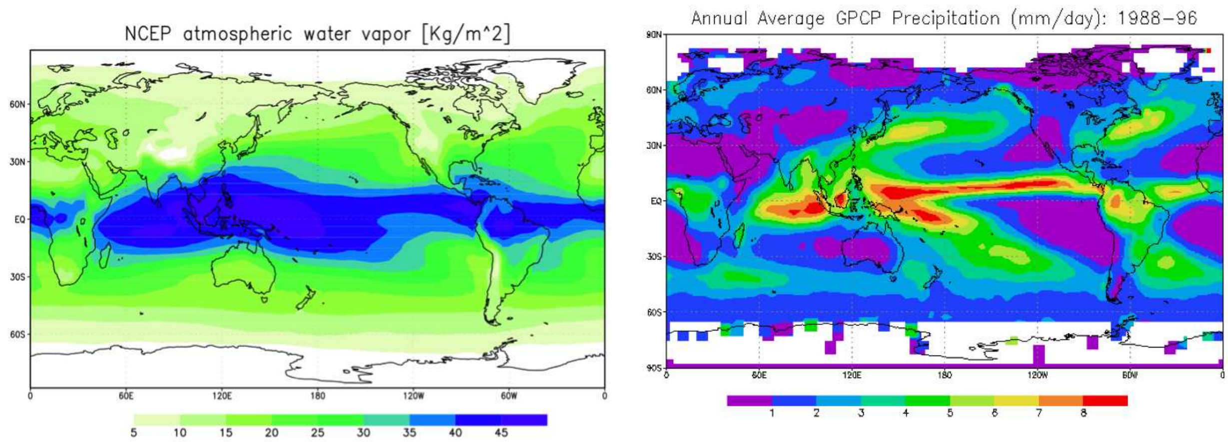

Figure 4.20: Left: Mean atmospheric water vapour. Right: Mean precipitation. Note the similarity in the patterns suggests that rain is roughly proportional to the amount of water vapour in the air.

$|\vec{u}_*|$ = effective wind speed
v_{soil} = surface moisture [%]
q_{sat} = saturated humidity [kg/kg]
q_{atmos} = densithumidity [kg/kg]

Saturation

A closed sample of air over a plane water surface in equilibrium between condensation and evaporation is said to be saturated.

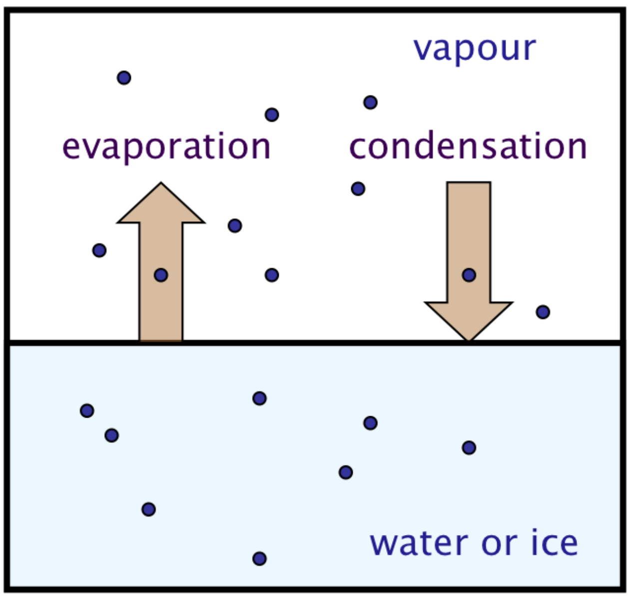

Figure 4.21: Sketch illustrating Saturation: Air over a moist surface is saturated if evaporation and condensation are in balance.

Water Vapour Mixing Ratio

Typically r and q range from 20 g/kg at low levels in the tropics to very low values at high latitudes or at high elevations.

If neither evaporation nor condensation take place, the mixing ratio of an air parcel is constant.

20 g/kg $\approx \frac{1}{2}$ small glass of water per 1 m^3
20 g/kg $\approx$ 10 litres per lecture room

Saturated water vapour: The amount of water vapour that the air can hold depends on the temperature; following from *Clausius-Clapeyron equation*:

$$q_{sat} = e^{\frac{-z_{topo}}{h_{atmos}}} \cdot 3.75 \cdot 10^{-3} \cdot e^{17.1 \frac{T_{surf}-273.15}{T_{surf}-38.98}} \tag{4.14}$$

This equation also considers that the surface pressure decrease with altitude. The derivation of this equation (without the topography effect) and its solution is given in most texts on thermodynamics.

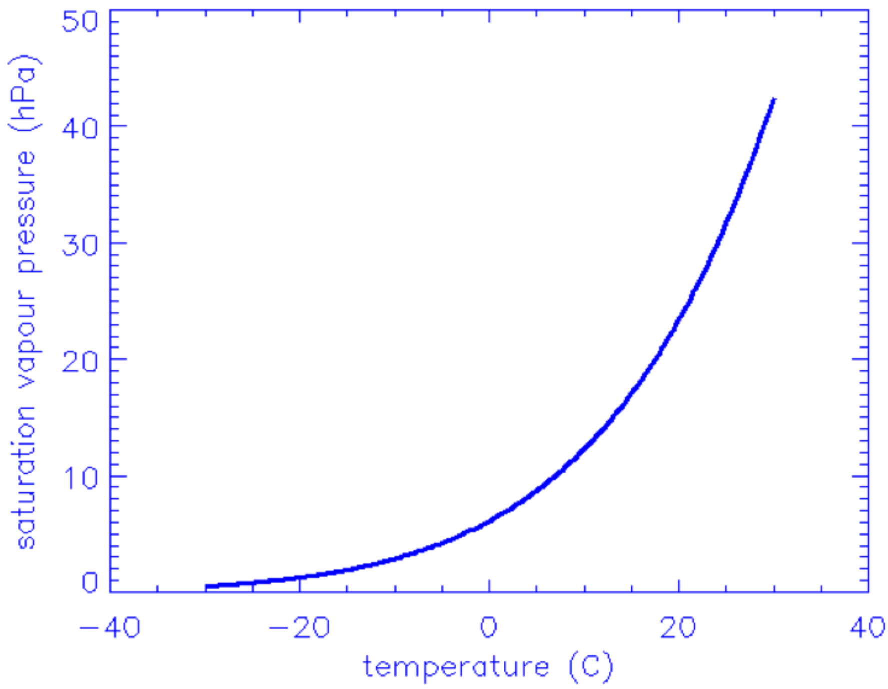

Figure 4.22: Saturation vapour pressure as function of temperature. The Clausius-Clapeyron relation.

Note: if $dT = 3^oC \rightarrow q_{sat}$ changes by about 20%. Assuming relative humidity is not changing, global warming will most likely cause a significant change in q_{atmos}.

Over all we find that Evaporation is strong if:

- Winds are strong.
- Air is dry and warm.
- Surface is wet.

Surface Humidity vs. Total water vapour in air column: In the emissivity function for the thermal radiation we need to know the total amount of water vapour in the air column, H_2O. In the above hydrological cycle we discussed terms for the surface humidity change, q_{atmos}.
In the GREB model we make the simple approximation:

$$H_2O_{vapour} = e^{-z_{topo}} h_{atmos} \cdot r_{H_2O} \cdot q_{atmos} = e^{-z_{topo}} h_{atmos} \cdot 2.67 \cdot 10^3 kg/m^2 \cdot q_{atmos} \tag{4.15}$$

The total amount of water vapour in an air column is proportional to the surface humidity and we assume some altitude effect, that reduces the total amount of water vapour in the air column for a given surface humidity.

Latent heat: F_{latent}

$F_{latent} = -L \cdot r_{H_2O} \cdot \Delta q_{eva}$

$L = 2.3 \cdot 10^6 J/kg$ = latent heat of condensation
$c_{H_2O} \approx 4000 J/kg/K$ = specific heat of water

Note: this is a lot of heat if compared to the specific heat of water. Condensing water vapour to water (raining) releases more heat than is required to heat water by 500 K!!

Latent heating implications for deep tropical convection

Conclusion: Deep tropical convection cannot be dry adiabatic.

$\gamma_{surf} \frac{dT_{surf}}{dt} = F_{solar} + F_{thermal} + F_{latent} + \ldots$

$\gamma_{atmos} \frac{dTatmos}{dt} = F_{thermal} + F_{atmos-latent} + \ldots$

$\frac{dq_{surf}}{dt} = \Delta q_{eva} + \Delta q_{precip} + \ldots$

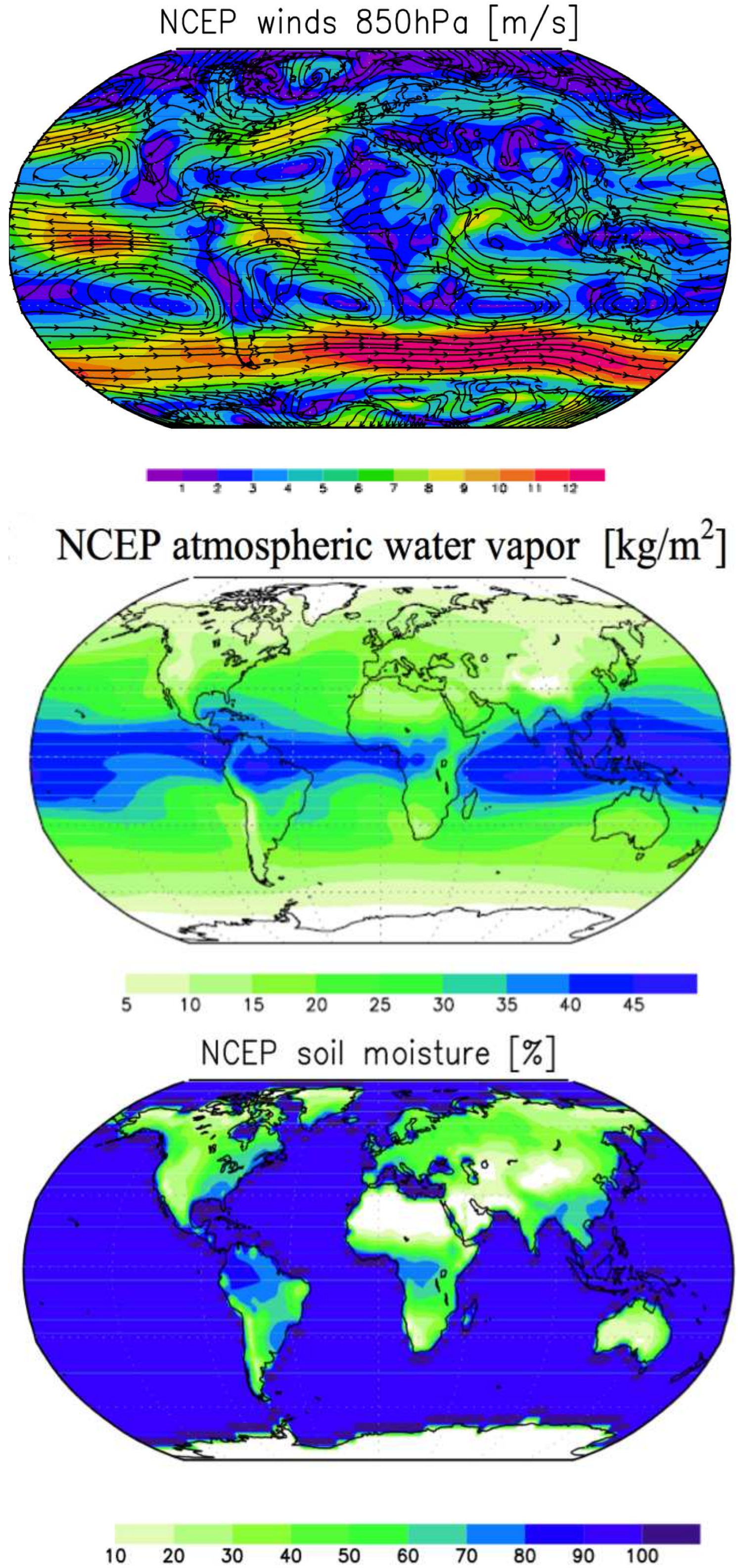

Figure 4.23: Climatic fields important for evaporation: Surface winds, soil moisture and atmospheric near surface humidity.

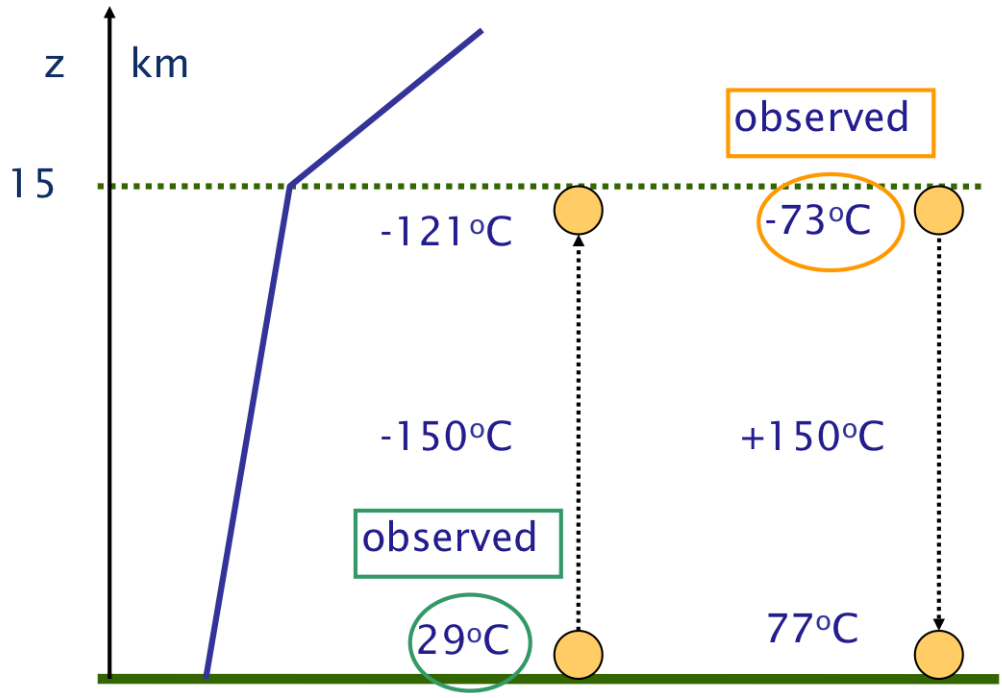

Figure 4.24: Illustration of tropical vertical temperature gradients. When air rises in convective thunderstorms, it looses all of its water vapour by raining or building clouds. This releases latent heat and heats the tropical atmosphere by about $+50^oC$. The vertical temperature gradient is much weaker than you would expect from the adiabatic cooling by expanding into the lower pressure high levels.

4.1.5 Sensible heat

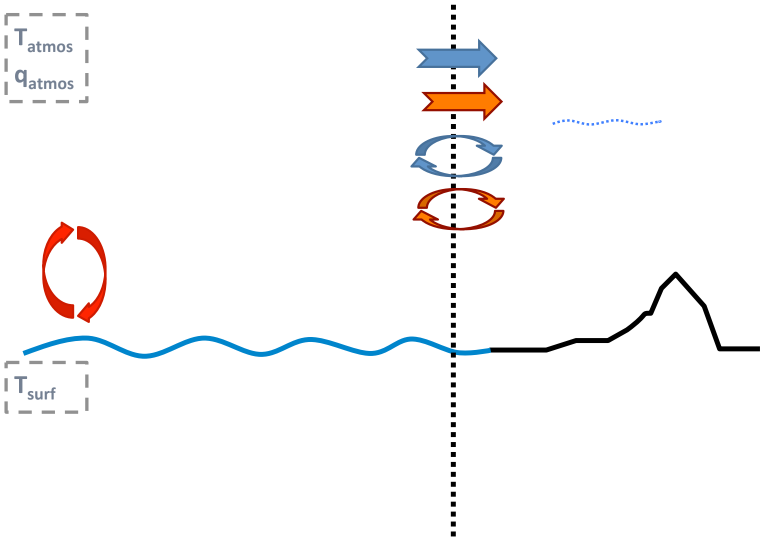

Figure 4.25: Sketch GREB model process: Sensible heating and transport of heat and moisture in the atmosphere.

Sensible heat and circulation

Simplification:

Assume heat transport only in atmosphere (not in ocean, not in surface layer)

Sensible heat flux from atmosphere:

$$F_{sense} = C_{A-S} \cdot (T_{atmos} - T_{surf})$$
$$C_{A-S} = 22.5 \frac{W}{m^2} \frac{1}{K}$$

The turbulent heat exchange between surface and atmosphere is estimated by a ***Newtonian damping***. The stronger the temperature difference the stronger the heat exchange.

$$F_{sense} - C_{O-S} \cdot (T_{ocean} - T_{surf})$$
$$C_{A-S} = 5 \frac{W}{m^2} \frac{1}{K}$$

The turbulent heat exchange with subsurface ocean is much weaker due to the much weaker turbulence in the subsurface ocean.

Atmospheric heat transport

Simplification of transport:

- Assume a mean transport (advection)
- Assume a turbulent isotropic diffusion

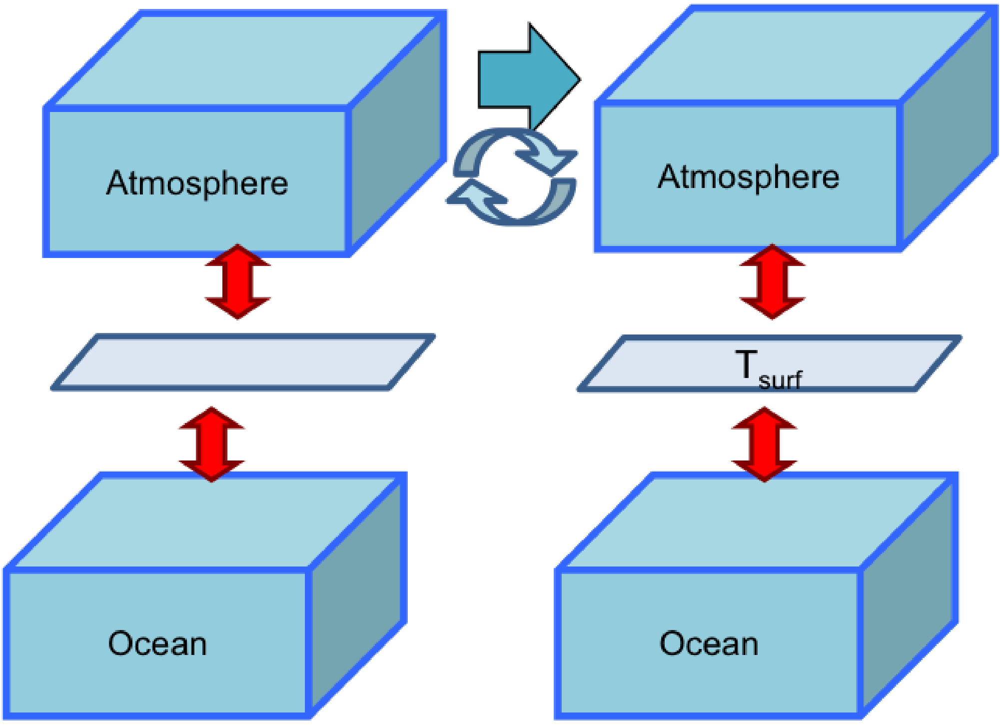

Figure 4.26: Sketch of the GREB sensible heat exchange. The surface exchanges sensible heat with the atmosphere and subsurface oceans. The heat is transported horizontally only in the atmospheric layer.

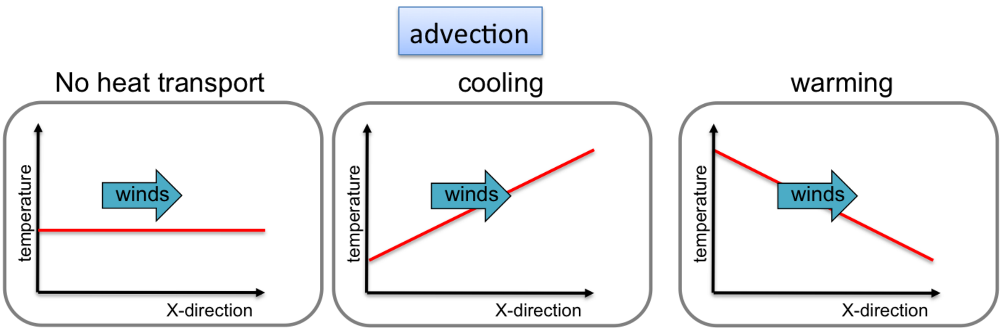

Figure 4.27: Advection with the mean winds transports heat across temperature gradients.

$F_{adv} \propto -\vec{u} \cdot \vec{\nabla} T$

Heat transport by advection is proportional to the scalar product between wind-vector and the direction of the temperature gradient.

$\Rightarrow \frac{F_{adv}}{\gamma_{atmos}} = -\vec{u} \cdot \vec{\nabla} T$

The temperature tendencies by advection does not depend on the heat capacity, but the heat flux does.

Isotropic diffusion (mixing by turbulent winds) heats or cools regions with extremes (minimas or maximas).

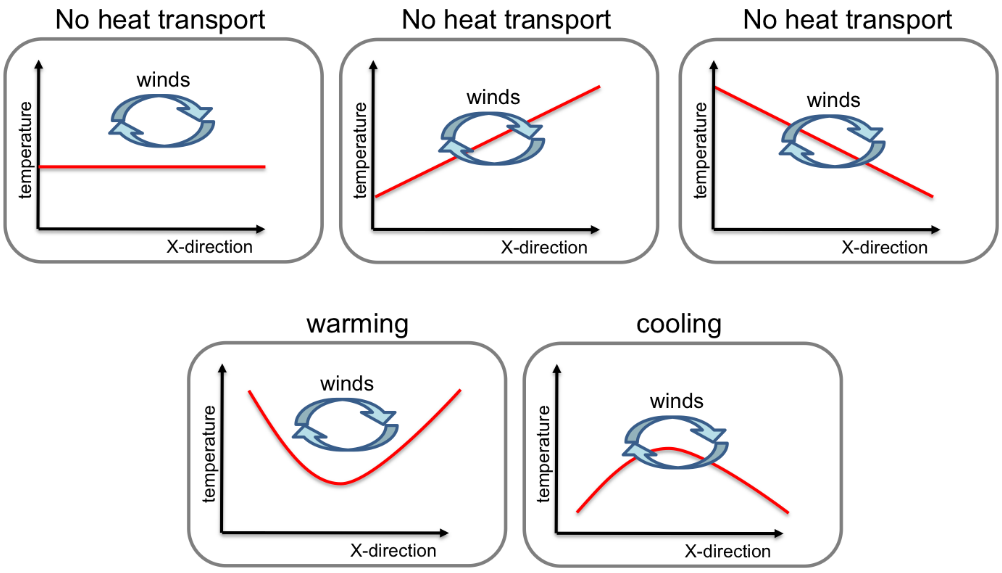

Figure 4.28: Isotropic diffusion

Diffusion transports heat away or to regions which are warmer or cooler than the neighbourhood.

$\frac{F_{diffuse}}{\gamma_{atmos}} = \kappa \cdot \nabla^2 T$

Heat transport by isotropic diffusion is strong if the 2nd derivative of the temperature field is strong and the turbulence of the winds (kappa) is strong.

$\kappa = 2 \cdot 10^5 \frac{m^2}{2}$

This defines the strength of turbulent winds (weather). It should erase any temperature maxima/minima in the atmosphere within days.

4.1.6 Subsurface Ocean

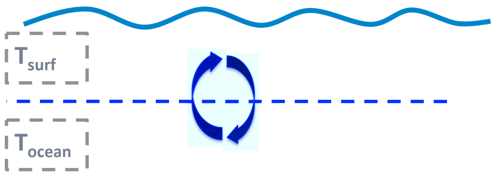

Figure 4.29: Sketch GREB model process subsurface heat exchange.

Ocean heat uptake: The ocean takes up a lot of heat. The heat that goes in to the subsurface ocean is not just a function of the surface temperature change, but can be independent of the surface temperature. We therefore need to simulate a subsurface ocean temperature, T_{ocean}.

Simplification:

- No lateral heat transport in ocean. We assume all heat is transported in the atmosphere.
- Sensible heat flux (Newtonian damping).
- Entrainment by changes in mixed layer depth (MLD). The seasonal cycle in the thickness of the surface layer (mixed layer depth) does most of the heat exchange between the surface layer and the subsurface ocean.
- Effective ocean heat capacity is porportional to MLD. Regions that have a deeper surface layer (mixed layer depth) will also mix deeper into the ocean.

Sensible heat flux from deeper ocean:

$$F_{O_{sense}} = C_{O-S} \cdot (T_{ocean} - T_{surf})$$

$$C_{A-S} = 5 \frac{W}{m^2} \frac{1}{K}$$

$$F_{ocean} = F_{O_{sense}} + \mathbf{\Delta T_{entrain}}$$

Is strong if the seasonal cycle of the mixed layer depth is strong.

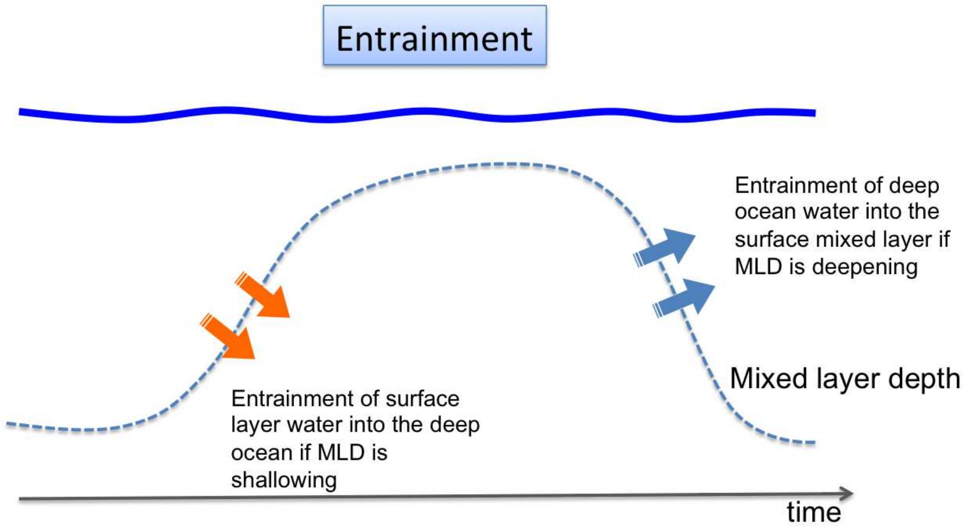

Figure 4.30: Entrainment: the seasonal cycle of the surface mixed layer depth causes most of the heat exchange between the surface and the subsurface oceans.

Subsurface ocean temperature tendencies

$$\gamma_{ocean}\frac{dT_{ocean}}{dt} = \Delta T_{O_{entrain}} - F_{O_{sense}} \tag{4.16}$$

Ocean mixed layer depth

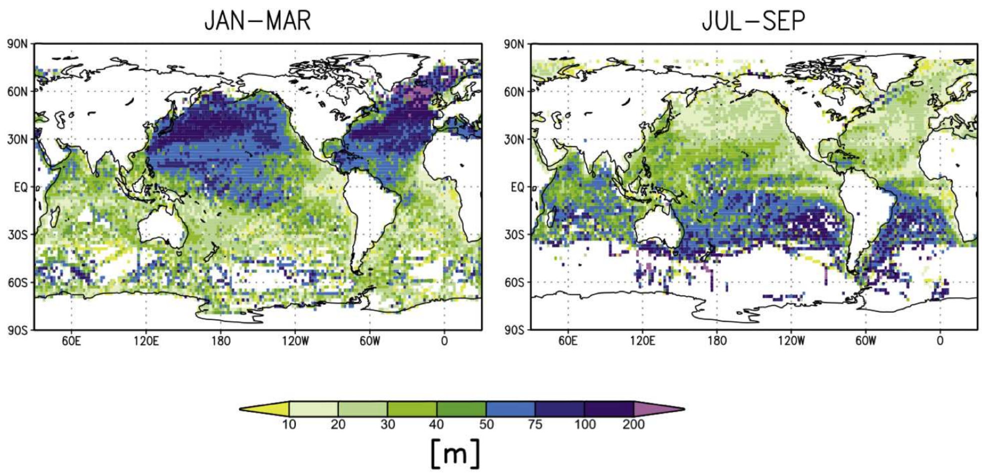

Figure 4.31: Ocean mixed layer depth. Regions with deep mixed layers are also regions in which a lot of heat is exchanged with the deeper oceans.

Heat capacity

Atmosphere	$\gamma_{atmos} = 5 \cdot 10^6 \cdot J/m^2/K$	About 5000m air column
Land	$\gamma_{surf} = 4 \cdot 10^6 \cdot J/m^2/K$	About 2m soil column
Ocean surface	$\gamma_{surf} \approx 200 \cdot 10^6 \cdot J/m^2/K$	In average 50m water column; large regional and seasonal differences depending on mixed layer depth
Subsurface ocean	$\gamma_{ocean} \approx 1000 \cdot 10^6 \cdot J/m^2/K$	About three times the max. mixed layer depth; strong regional differences. Not the whole deep ocean is mixed.
Sea ice	$\gamma_{surf} = 4 \cdot 10^6 \cdot J/m^2/K$	As for ice-albedo feedback, some ocean points are sometimes ice covered. Sea ice insulates the ocean from the atmosphere very well. → heat capacity over ocean is function of temeprature for below freezing values.

4.1.7 Sea Ice

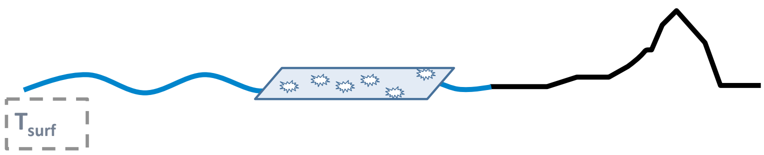

Figure 4.32: Sketch GREB model process sea ice.

Sea ice heat capacity

Sea ice: $\Rightarrow \gamma_{surf} \approx$ 2m water column

Open ocean: $\Rightarrow \gamma_{surf} \approx$ 20 - 300m water column

$\Rightarrow \gamma_{surf} = \gamma_{surf}(T_{surf})$

Sea ice has a special feedback on the climate, by changing the effective heat capacity.

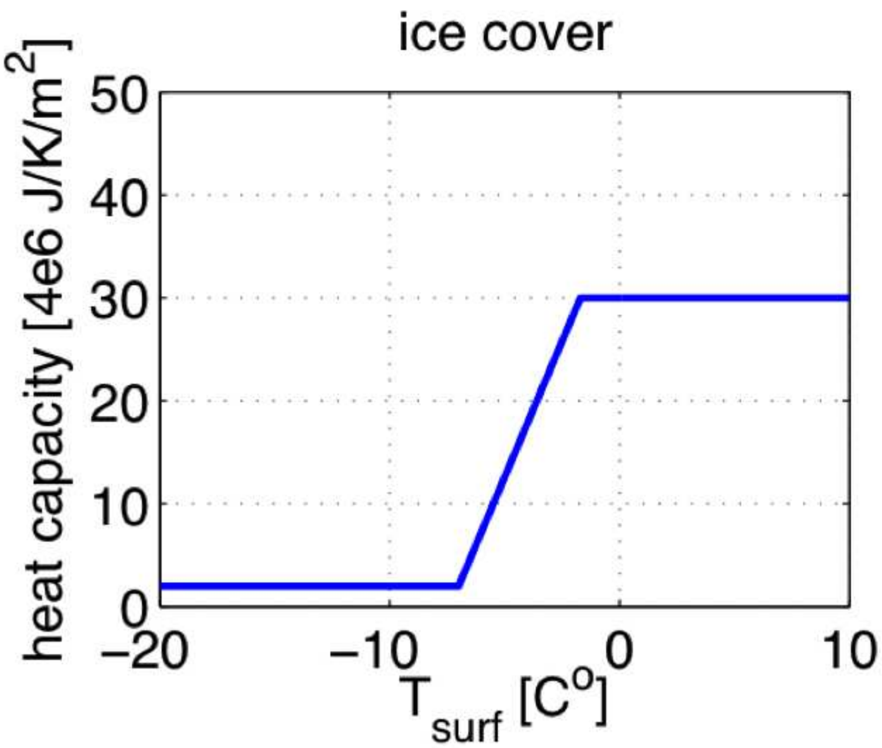

Figure 4.33: Sea ice heat capacity function for the GREB model.

4.1.8 Estimating the equilibrium climate

Response time

How long does it take for the climate to respond to forcing?

$$\gamma \frac{dT}{dt} = C_f \cdot T + Q \quad \begin{array}{l} C_f \approx 1 \frac{W}{m^2} \frac{1}{K} \\ Q \approx 4 \frac{W}{m^2} \end{array}$$

It takes about 20-30 years for the surface ocean to respond, but the deep ocean will take much longer ($\tilde{1}$000 years).

Model integration

A climate model is started with some initial condition (observed climate state) and the tendencies are added (integrated) with small time steps (Δt). If the model is not perfect (no model is perfect) it will drift away from the observed climate. The new equilibrium climate state will be different from the observed.

Note, the tendencies are different in each time step as the depend on the climate state of the previous time step.

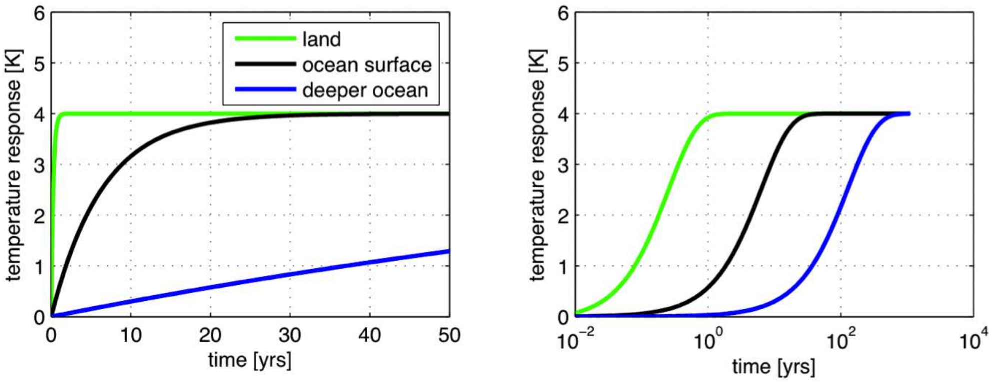

Figure 4.34: Response time to a fixed forcing in a simple linear response model.

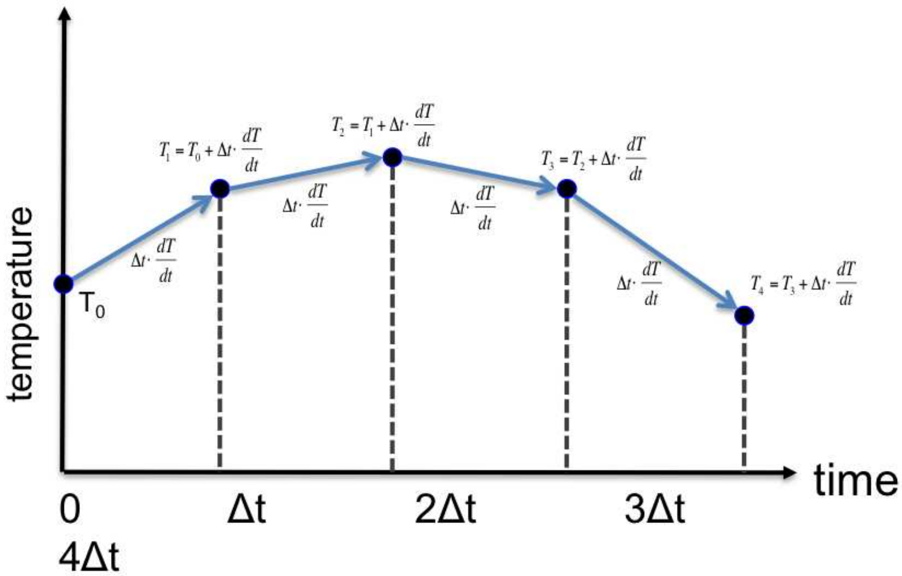

Figure 4.35: Numerical integration of tendency equations. Complex systems that can not be solved analytically, have to be integrated by small time steps Δt for the initial value T_0 to the future.

$$\gamma_{atmos}\frac{dT}{dt} = F(t, T, q, \dots)$$
$$\int\limits_{T_{start}}^{T_{end}} dT = \int\limits_{t=0}^{t_{end}} F(t, T, q, \dots)dt$$
$$T_{end} - T_{start} - \frac{1}{\gamma}\int\limits_{t=0}^{t_{end}} F(t, T, q, \dots)dt$$

4.1.9 Numerical Simulations with the GREB model

GREB model correction

The GREB model is too simple to produce a realistic climate state. If we integrate the model from an observed climate initial condition it will drift into a climate state that is very different (10K or more) from the observed.

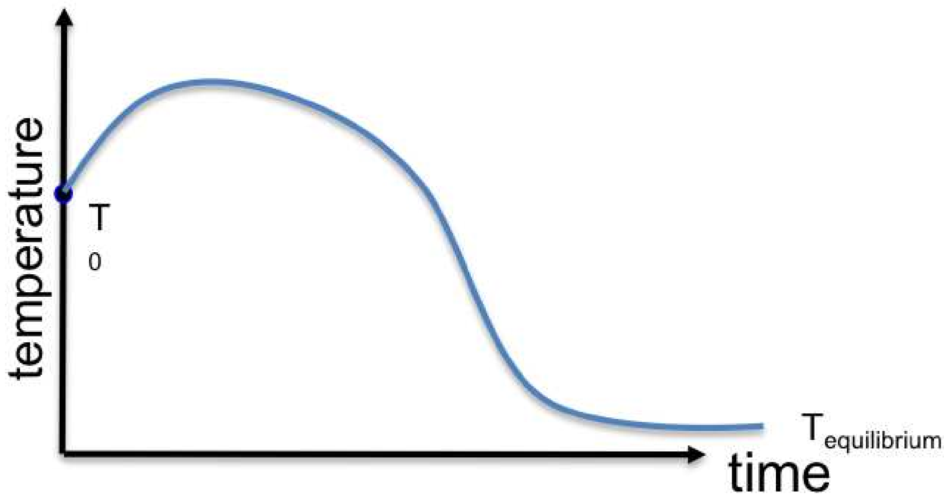

Figure 4.36: A imperfect climate model (all models are imperfect) will drift from an observed mean (equilibrium maybe?) climate state to another equilibrium climate state different from the observed if it is integrated over time from the mean obsevred climate state.

Note: The climate mean state has a range from -50° C to 40° C.So it is a $\Delta\ T \approx 100$ ° C for largely different forcings at different regions and seasons. Climate change is a $\Delta\ T \approx 5°$ C for a small forcing, which is much smaller than the range in mean climate state. So a model which may have a 10-20% error in the tendencies will lead to a large error in the mean climate, but will have an 'OK' error for climate change.

Flux corrections

To keep GREB close to the observed climate state we need to introduce artificial heat fluxes to correct the tendencies. These correction terms in the tendency equations will force the GREB model to produce exactly the observed mean T_{surf} and mean q_{surf}.

GREB boundary conditions: *external*

IPCC models will have similar boundary condition.

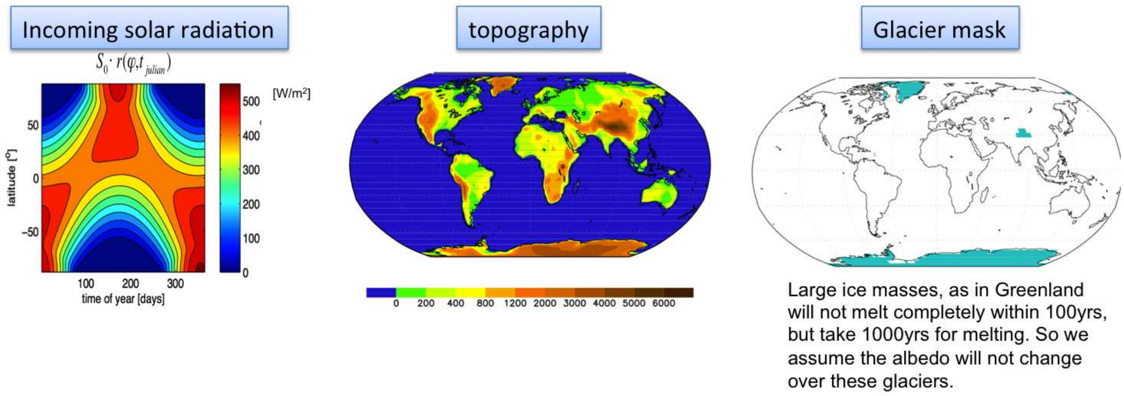

Figure 4.37: External boundary conditions for the GREB model.

Additionally there are some physical constants for the climate system: e.g. surface air pressure, CO_2, concentration, etc.

GREB boundary conditions: *internal*

IPCC models will simulate these internal boundary conditions. So for an IPCC model these are not boundary conditions, but are computed by the state of the system with prognostic equations.

All given with seasonally changing climatology.

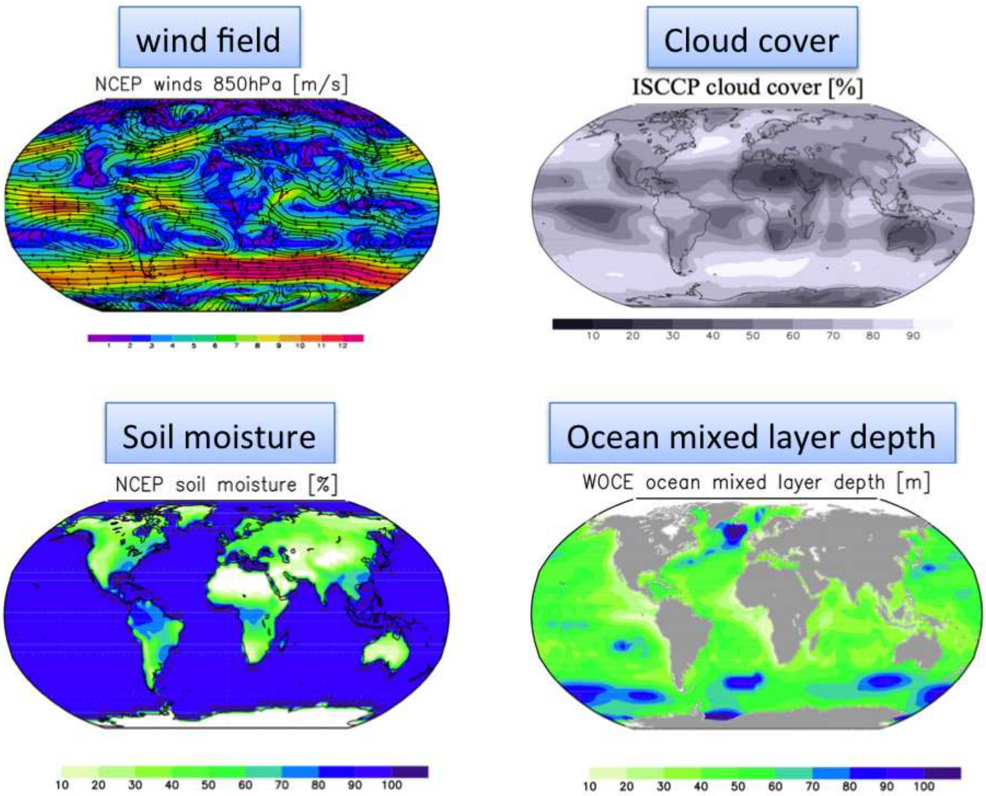

Figure 4.38: Internal boundary conditions for the GREB model. These are not really boundary conditions, but are parts of the climate system that the GREB model assumes to be given. In the real world these are variable and will change over time.

GREB constraints

IPCC models will simulate these variables without artificial constraints.

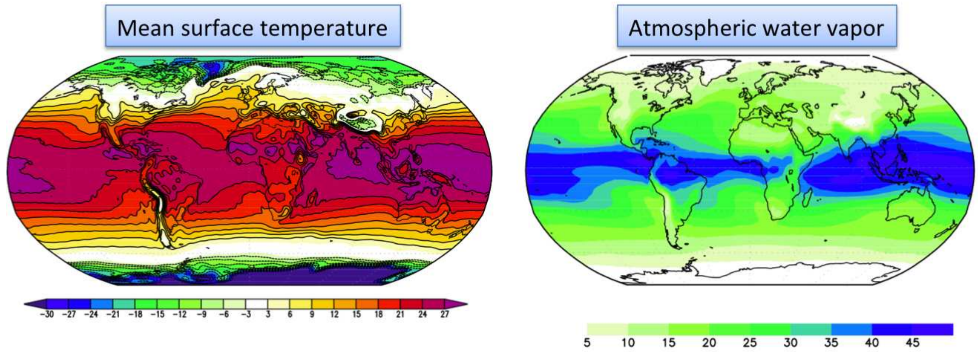

Figure 4.39: ExtConstraints for the GREB model. We artificially force the GREB model to have the mean observed surface temperature and atmospheric surface humidity by flux correction terms.

Enforced in GREB by flux correction terms.

GREB response

Climate response to CO_2 forcing scenario IPCC A1B:

time intervals: [2070-2100] - [1970-2000]

	IPCC	**GREB**	**Comment**
Global mean	2.6K	2.7K	This is somewhat expected as we have fitted our emissivity function by the IPCC results
Land warms more than oceans	60% more	30% more	GREB is not warming enough over land
Polar amplification	By factor 2	By factor 1.3	GREB is not warming enough over Arctic sea
Northern Hemisphere warms more than the Southern	Yes	Yes	GREB inter-hemispheric contrast is weaker
Cold season warms more	Yes	Yes	-

The GREB model is able to simulate the main IPCC T_{surf} response structures.

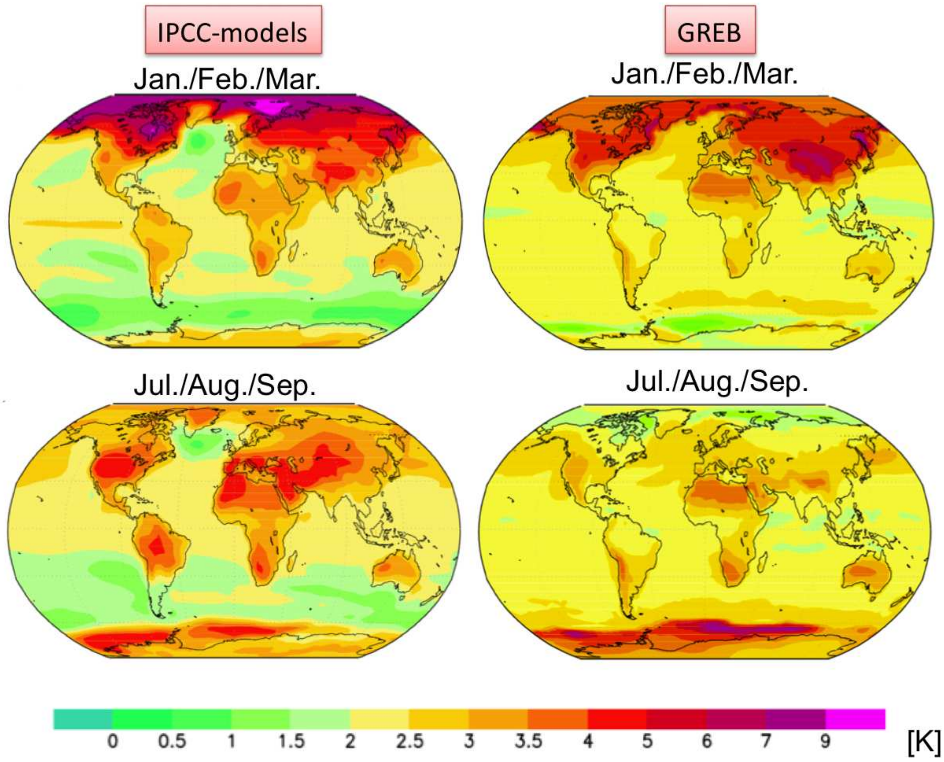

Figure 4.40: Comparison of the surface temperature response to CO_2-forcing IPCC models vs. GREB. The main features are well simulated by the GREB model. Note, that we do not know what the 'real' response is, yet.

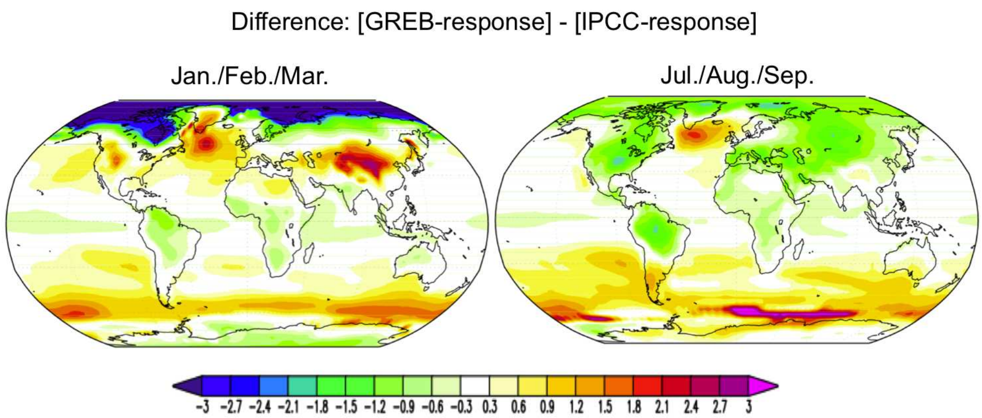

Figure 4.41: Difference in the surface temperature response to CO_2-forcing IPCC models vs. GREB.

- Southern ocean warms too much: IPCC models show strong heat uptake in the Southern Ocean. Unclear if this is the problem in GREB.
- Northern North Atlantic warms too much : IPCC models show a significant reduction of ocean transport (Gulf Stream), which is generally assumed to cause the weak warming in the North Atlantic.

IPCC warming response

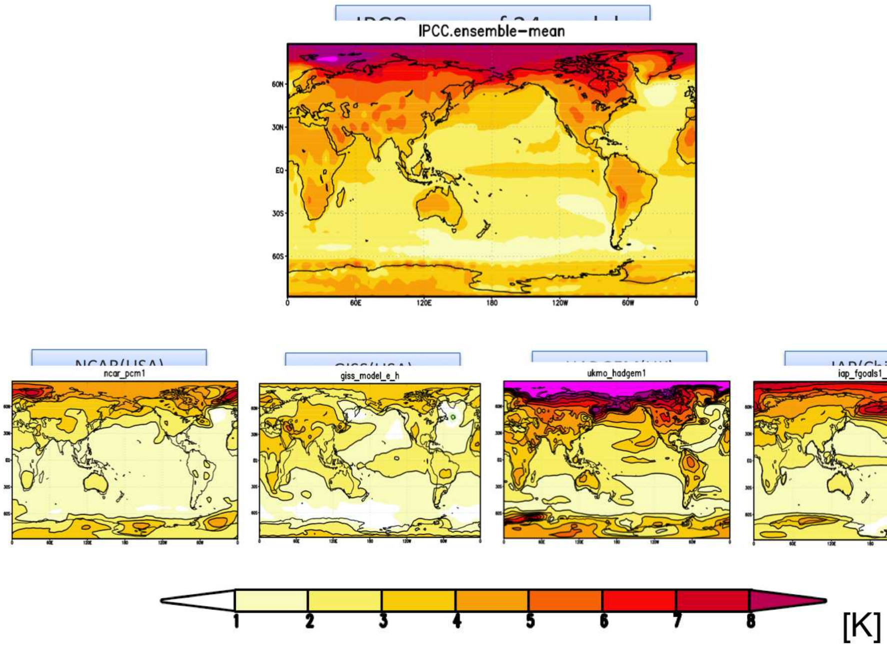

Figure 4.42: Upper: IPCC ensemble mmean response to CO_2-forcing of 24 IPCC models. Lower: 4 example models. We can see that different models predict quite different warming patterns and global mean amplitudes.

GREB response 'skill'

The GREB T_{surf} response is within the uncertainty of the 24 IPCC model predictions. Note, nothing is said about whether or not this response is similar to what the real world is doing.

GREB response in other variables

Response in relative humidity:
Mean values range from 20-80%

- mostly ± 3%, which is basically no change
- drying over land
- more humid over oceans

IPCC models: Similar, with mostly no significant change in humidity.

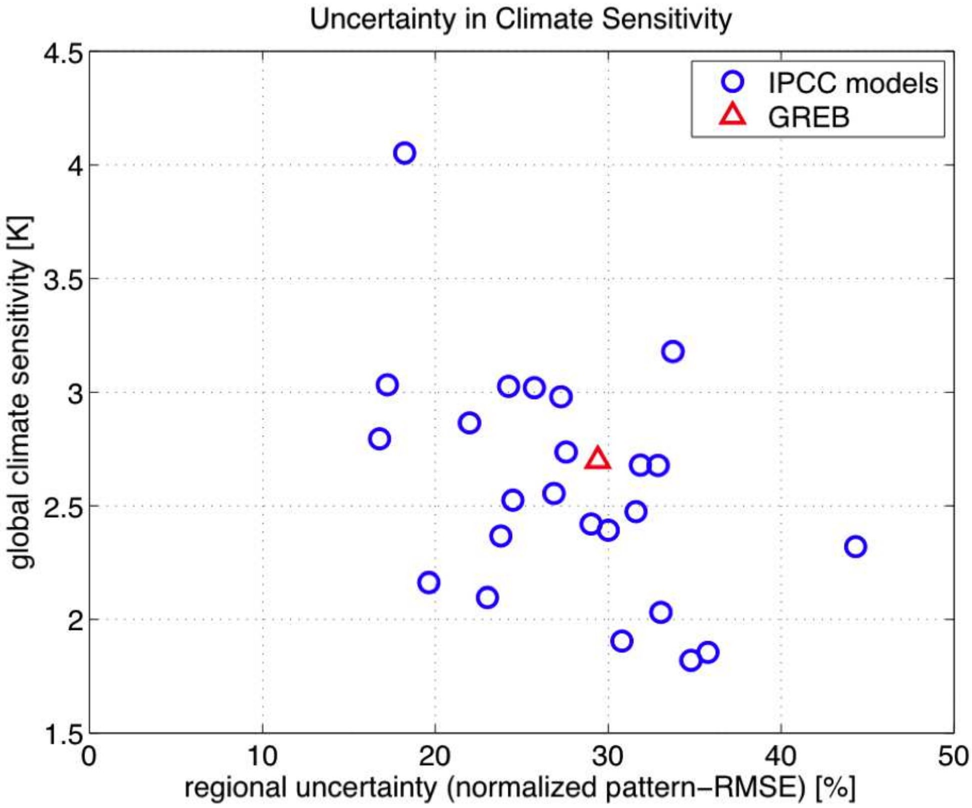

Figure 4.43: Spread in the model response. y-axis: The global mean warming. x-axis: similarity to the IPPC ensemble mean warming. 0 means the same warming pattern and 50% means in average a 50% different response amplitude.

Response in precipitation:

- increasing everywhere by 10-20%
- more strongly over dry regions
- GREB can only have increase in rain if q_{surf} increases - precipitation : $\Delta Q_{precip} = 0.1\frac{1}{day}q_{surf}$

IPCC models: Mostly increasing by about 5-10%, but dry regions will mostly have a decrease in precipitation. Some models have quite significant decrease in precipitation.

GREB response in other variables

Response in T_{atmos}:

- T_{atmos} warms more than T_{surf}
- most strongly in the warm and wet regions
- this is caused by increased precipitation and the associated latent warming in the atmosphere and latent cooling at the surface

IPCC models: Similar, but more vertical structure of cause.

Response in ocean heat uptake:

- mostly in high latitudes - Southern Ocean, Atlantic and Pacific
- caused by deeper mixed layer depth in these regions

IPCC models: Similar in pattern and strength, but stronger over Southern Ocean.

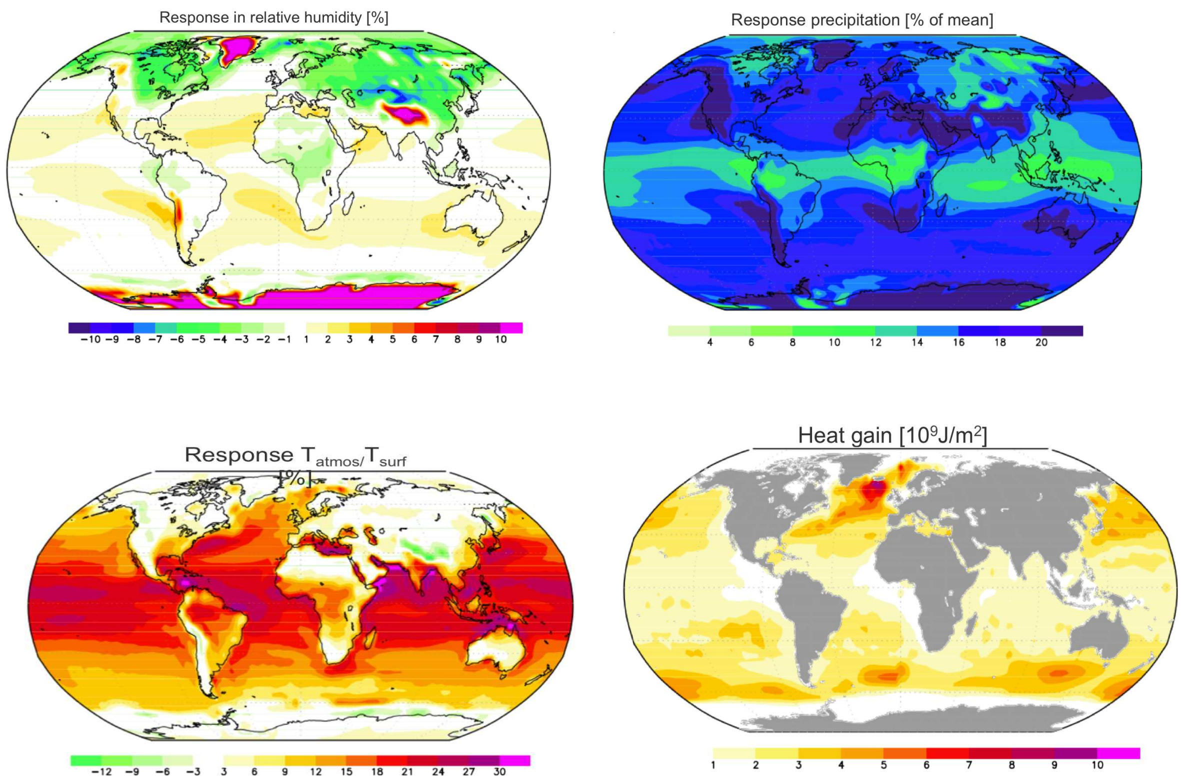

Figure 4.44: GREB response in other variables: Upper left: relative humidity in [%]. Upper right: precipitation in percentage of the mean precipitation. Lower left: Ratio of $\Delta T_{atmos}/\Delta T_{surf}$. Lower right: Oceans heat up take in $[10^9 J/m^2]$.

GREB response precipitation:

Response precipitation [% of mean]

4.1.10 A short summary of the GREB model

4.2 Conceptual deconstruction of climate change

In the following we will use the GREB model to deconstruct the surface temperature response to increases in CO_2 concentrations. This will help us to understand how the different processes interact to cause the main structures of the surface temperature response.

We do this by sensitivity experiments with the GREB model in which we turn some processes of the model "*off*". We will discuss 10 experiments that start with the simplest sensitivity experiments and ends with the complete GREB model. The experiments [1] to [4] focus on the local response to the direct CO_2 forcing without any climate feedbacks. Before we start the discussion of the feedbacks we discuss the role of heat advection in Exp. [5]. In the experiments [6] we address the ice/albedo and sea ice feedbacks and in the Exps. [7] to [9] we discuss the most important water vapor feedback. Finally we will discuss the ocean heat up take in Exp. [10].

4.2.1 The Direct Local Forcing Effect - No Feedbacks (Exp. [1] to [4])

We start the series of experiments with 4 experiments where most processes are '*turned off*'.

- No ocean heat uptake:
  - $\rightarrow F_{ocean} = 0$
  - $\rightarrow T_{ocean} =$ fixed
- No feedbacks:
  - $\rightarrow$ albedo = fixed
  - $\rightarrow q_{surf} =$ fixed
  - $\rightarrow$ No latent heating
- No circulation:
  - $\rightarrow$ advection = 0
  - $\rightarrow$ diffusion = 0

With the circulation turned '*off*' we can discuss the response at each location independent of each other. Thus, effectively, each grid point of the GREB model is now responding independent of the others.

The four GREB Model tendencies equations are:

$$\gamma_{atmos}\frac{dT_{atmos}}{dt} = F_{thermal} + F_{atmos-latent} - F_{sense} + \gamma_{atmos}(\kappa \cdot \nabla^2 T - \tilde{u} \cdot \tilde{\nabla} T)$$

$$\gamma_{surf}\frac{dT_{surf}}{dt} = F_{solar} + F_{thermal} + F_{latent} + F_{sense} + F_{ocean} + F_{correct}$$

$$\gamma_{ocean}\frac{dT_{ocean}}{dt} = \Delta T_{O_{entrain}} - F_{O_{sense}} - F_{O_{correct}}$$

$$\frac{dq_{surf}}{dt} = \Delta q_{eva} + \Delta q_{precip} + (\kappa \cdot \nabla^2 q_{surf} - \tilde{u}\tilde{\nabla} q_{surf}) + \Delta q_{correct}$$

With the processes turned '*off*' these equations reduce substantially:

$$\gamma_{atmos}\frac{dT_{atmos}}{dt} = F_{thermal} - F_{sense} \tag{4.17}$$

$$\gamma_{surf}\frac{dT_{surf}}{dt} = F_{solar} + F_{thermal} + F_{sense} + F_{correct} \tag{4.18}$$

To maintain the same reference mean T_{surf} we replace missing processes by additional flux corrections ($F_{correct}$).

Experiment [1]: The Pure Radiation Balance - No Regional Difference in Greenhouse Effect

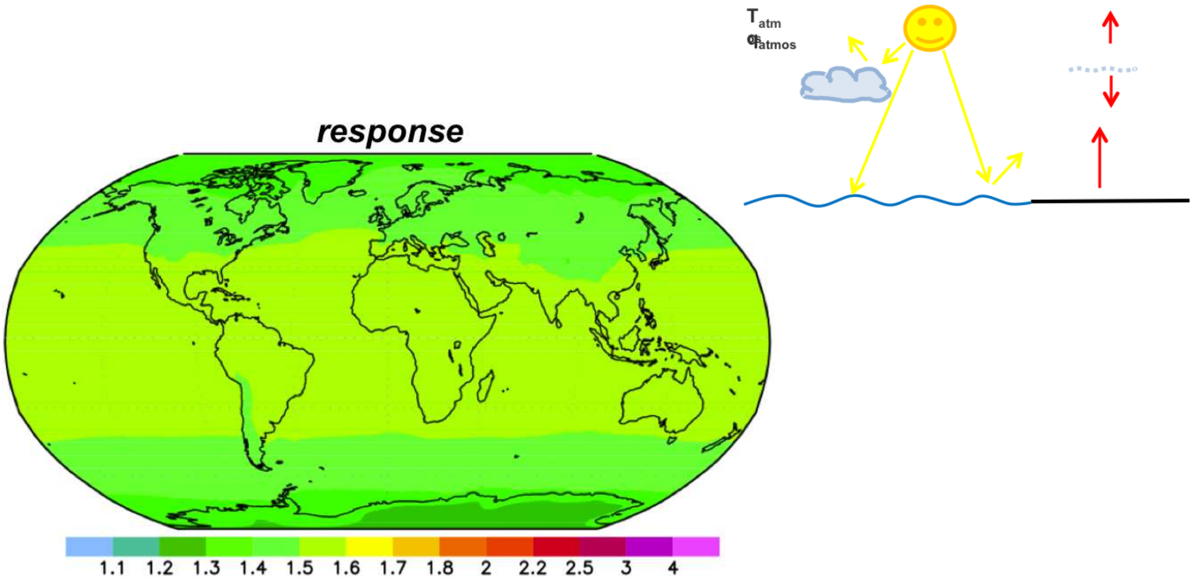

Figure 4.45: Exp. [1] The Pure Radiation Balance: No feedback, No topography or transport of heat; homogenous cloud cover (0.7) and water vapour. The response pattern is only caused by differences in T_{surf}. Colour bar is the response in T_{surf} to a doubling of CO_2 concentrations in $[^oK]$. Note that the shading interval is non-linear.

For Exp. [1] we further set all boundary conditions that affect the thermal radiation to be globally uniform:

- No topography
- globally uniform clouds ($a_{cloud} = 0.7$)
- globally uniform water vapour ($q_{surf} = 0.0052$ and thus $H_2O = 14Kg/m^2$)

Most terms in the surface temperature tendencies equation are now constant (not changing over time other then the seasonal cycle). We can therefor now simplified our surface temperature tendencies equation:

$$\gamma_{surf} \frac{dT_{surf}}{dt} \approx F_{thermal} + F_{constant} \tag{4.19}$$

So only the $F_{thermal}$ term is depending on the climate and therefor changing over time. This strong simplification of the GREB model allows us to estimate the surface temperature response to increases in CO_2 concentrations analytically.

We can approximate the $F_{thermal}$ term by a linearisation:

$$F_{thermal} \approx C_{thermal} \cdot T_{surf} + Q_{CO_2} \tag{4.20}$$

with $C_{thermal} = \frac{dF_{thermal}}{dT_{surf}}$ the linear feedback parameter as defined in eq.[2.50]. The CO_2 forcing is now only included in the forcing Q_{CO_2}.

In equilibrium, we have now:

$$C_{thermal} \cdot T_{surf} + Q_{CO_2} + F_{constant} = 0 \tag{4.21}$$

The equilibrium response in the surface temperature to increases in CO_2 concentrations, ΔT_{surf}, is the difference in this equation between $Q_{CO_2}(2xCO_2)$ and $Q_{CO_2}(control)$, ΔQ_{CO_2}:

$$C_{thermal} \cdot \Delta T_{surf} + \Delta Q_{CO_2} = 0 \tag{4.22}$$

with $\Delta T_{surf} = T_{surf}(2xCO_2) - T_{surf}(control)$ and $\Delta Q_{CO_2} = Q_{CO_2}(2xCO_2) - Q_{CO_2}(control)$. Thus the equilibrium response is

$$\Delta T_{surf} = \frac{-\Delta Q_{CO_2}}{C_{thermal}} \tag{4.23}$$

We can estimate the linear feedback parameter $C_{thermal}$ by the thermal radiation term:

$$F_{thermal} = -\sigma T_{surf}^4 + \epsilon \sigma T_{atmos}^4 \tag{4.24}$$

The atmospheric temperature $T_{atmos} \approx 0.84 * T_{surf}$. We can therefore replace T_{atmos}

$$F_{thermal} \approx -\sigma T_{surf}^4 + \epsilon \sigma 0.84^4 T_{surf}^4 \tag{4.25}$$

Thus

$$F_{thermal} \approx (0.5\epsilon - 1)\sigma T_{surf}^4 \tag{4.26}$$

So we get

$$C_{thermal} \approx 4(0.5\epsilon - 1)\sigma T_{surf}^3 \tag{4.27}$$

Note, that the negative feedback of the thermal radiation is stronger (more negative) if ϵ is smaller. From the emissivity function we can estimate the mean $\epsilon \approx 0.8$ for the given cloud cover and mean water vapour, see Fig. 4.18.
So for a mean $T_{surf} = 288^o K$ we get

$$C_{thermal} \approx -3.3 \frac{W}{m^2 K} \tag{4.28}$$

This is similar to the values we discussed in the energy balance sections of the Budyko and the zero order model. Next we have can estimate the ΔQ_{CO_2}

$$\Delta Q_{CO_2} = \Delta F_{thermal} = F_{thermal}(2xCO_2) - F_{thermal}(control) \tag{4.29}$$

As we only care about linear approximation, we can assume in this estimate that $T_{surf} \approx constant$. Using eq.[4.25] we get:

$$\Delta Q_{CO_2} \approx 0.5 \Delta\epsilon \sigma T_{surf}^4 \tag{4.30}$$

with $\Delta\epsilon = \epsilon(2xCO_2) - \epsilon(control)$. We can again estimate this as $\Delta\epsilon \approx 0.025$ for the given cloud cover and mean water vapour, see Fig. 4.18. So for a mean $T_{surf} = 288^o K$ we get

$$\Delta Q_{CO_2} \approx +4.8 \frac{W}{m^2} \tag{4.31}$$

So the doubling of the CO_2 concentration gives an initial forcing of $+4.8\frac{W}{m^2}$. This value can be compare to those estimates from the IPCC report, see Fig. 2.1.
We can now use eq. [4.23] to get the equilibrium response

$$\Delta T_{surf} \approx \frac{-0.5\Delta\epsilon\sigma T_{surf}^4}{4(0.5\epsilon - 1)\sigma T_{surf}^3} = \frac{0.5\Delta\epsilon}{4(1 - 0.5\epsilon)} T_{surf} = 0.2 \cdot \Delta\epsilon \cdot T_{surf} \tag{4.32}$$

The direct response to doubling of the CO2 concentration is a warming of about +1.5 degrees. It is proportional to the absolute temperature. So it is slightly larger in the tropics than in the polar regions, see Fig. 4.45.

Experiment [2]: The Role of Altitude on CO_2 Forcing

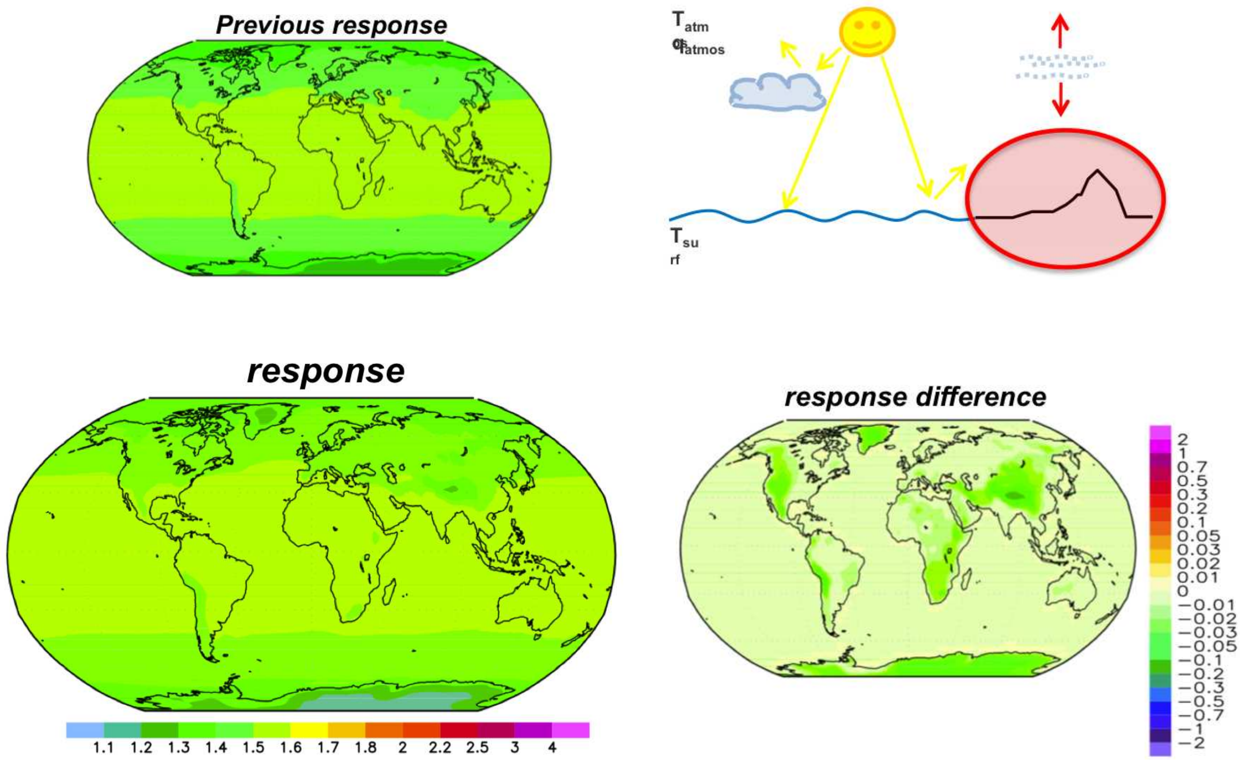

Figure 4.46: Exp. [2] Effect of topography: No feedback or transport of heat; homogenous cloud cover (0.7) and water vapour. Upper left: Response from the previous Exp. [1]. Lower left: The response pattern of Exp. [2]. Colour bar is the response in T_{surf} to a doubling of CO_2 concentrations in $[^oK]$. Lower right: difference of Exp. [2] minus previous Exp. [1]. It highlights the effect of high altitudes. Note that the shading intervals are non-linear.

- As [1] but with topography

In the emissivity function $\epsilon_{atmos} = F(CO_2^{topo}, H_2O, \text{cloud cover})$ the local CO_2 concentration, CO_2^{topo}, does depend on the topography:

$$CO_2^{topo} = CO_2 \cdot e^{\frac{z_{topo}}{h_{atmos}}} \tag{4.33}$$

Over high altitudes there is less atmosphere (lower pressure) and therefore less radiation effects. So the direct CO_2 effect is reduced.

Experiment [3]: ***Effect of the Mean Cloud Cover (No changes in clouds)***

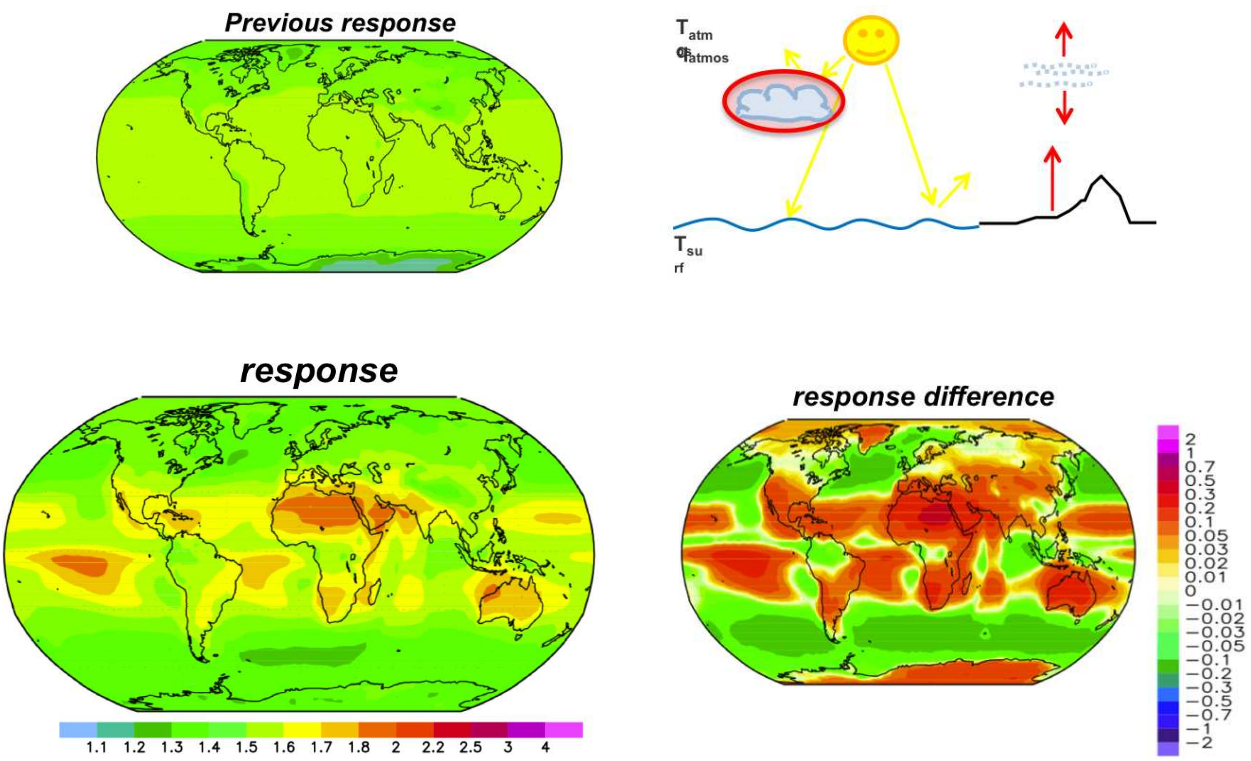

Figure 4.47: Exp. [3] Effect of mean cloud cover: No feedback or transport of heat; homogenous water vapour. The response difference highlights the effect of cloud cover. Details as in Fig. 4.46.

- As [2] but with true cloud climatology

$\epsilon_{atmos} = \frac{p_8 - a_{cloud}}{p_9}(\epsilon_{clear-sky} - p_{10}) + p_{10}$

- Cloud cover increases the mean ϵ
  - larger positive feedback by atmospheric thermal radiation (greenhouse effect)
  - more sensitive to external forcings
- Larger cloud cover dilutes the effect of the trace gasses ($\epsilon_{clear-sky}$) and therefore the CO_2 effect. So $\Delta\epsilon$ is smaller.

If we examine eq. [4.32] we see that a smaller $\Delta\epsilon$ would reduce the warming, but the larger mean ϵ would increase the warming. The later is less important for cloud cover. Overall cloudy regions will be less sensitive to CO_2 forcing than clear sky regions, but not as much as one may thing from the change in sensitivity to CO_2, because the increased mean ϵ is also increasing the positive feedback by *atmospheric thermal radiation*, see Fig. 4.48.

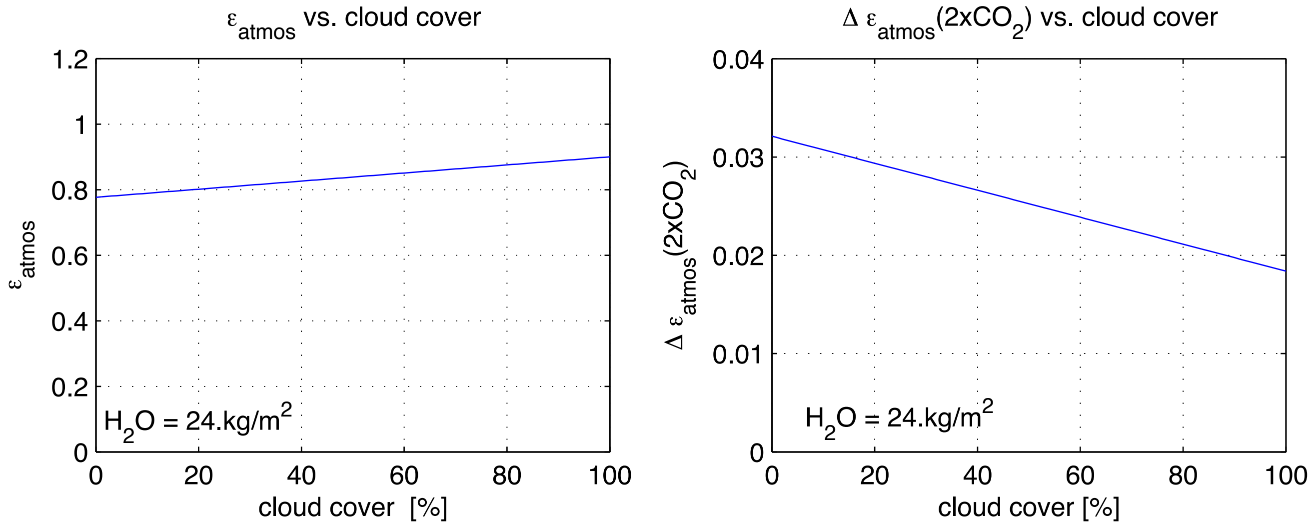

Figure 4.48: Sensitivity of the emissivity function to cloud cover.

Experiment [4]: ***Effect of the mean atmospheric water vapour (no feedbacks)***

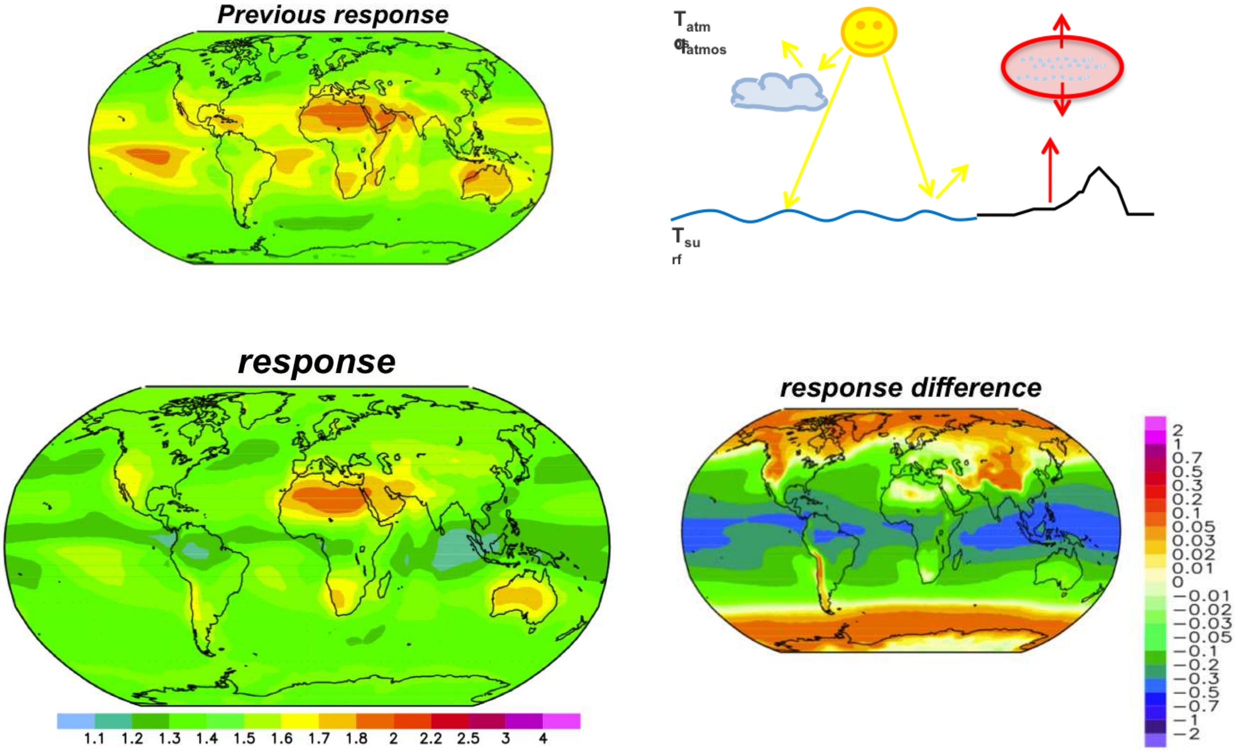

Figure 4.49: Exp. [4] Effect of the mean humidity: No feedback or transport of heat. The response difference highlights the effect of the mean humidity. Details as in Fig. 4.46.

- As Exp. [3] but with true humidity climatology
- Humid regions are less sensitive to CO_2 forcing due to the spectral absorption band overlap:

$$\varepsilon_{clear-sky} = \underbrace{p_4 \cdot \log\left[p_1 \cdot CO_2^{topo} + p_2 \cdot H_2O_{vapor} + p_3\right]}_{\text{CO}_2\text{ and H}_2\text{O absorb}} \; \underbrace{+p_5 \cdot \log\left[p_1 \cdot CO_2^{topo} + p_3\right]}_{\text{only CO}_2\text{ absorbs}} \; \underbrace{+p_6 \cdot \log\left[p_2 \cdot H_2O_{vapor} + p_3\right]}_{\text{only H}_2\text{O absorbs}} + p_7$$

The first term RHS is the log-function with both CO_2 and H_2O. If H_2O is low the CO_2 term has a bigger impact on the log-function and therefore on the emissivity. So a change in CO_2 leads to a bigger $\Delta\epsilon$. Further, if H_2O is low the overall emissivity ϵ is also low. If we examine eq. [4.32] we see that a smaller ϵ would increase the sensitivity and thus increase the warming. So lower H_2O increases the sensitivity by both the bigger $\Delta\epsilon$ and the smaller mean ϵ, see Fig. 4.50.

Summary of the direct local CO_2 forcing

- differences in cloud cover, humidity and topography cause difference in the local response to CO_2 forcing
- the strongest response is over the warm dry deserts

- the weakest response is over the warm humid oceans
- the global mean response is 1.5 ° C

Experiment [5]: ***Effect of atmospheric heat transport (no feedbacks)***

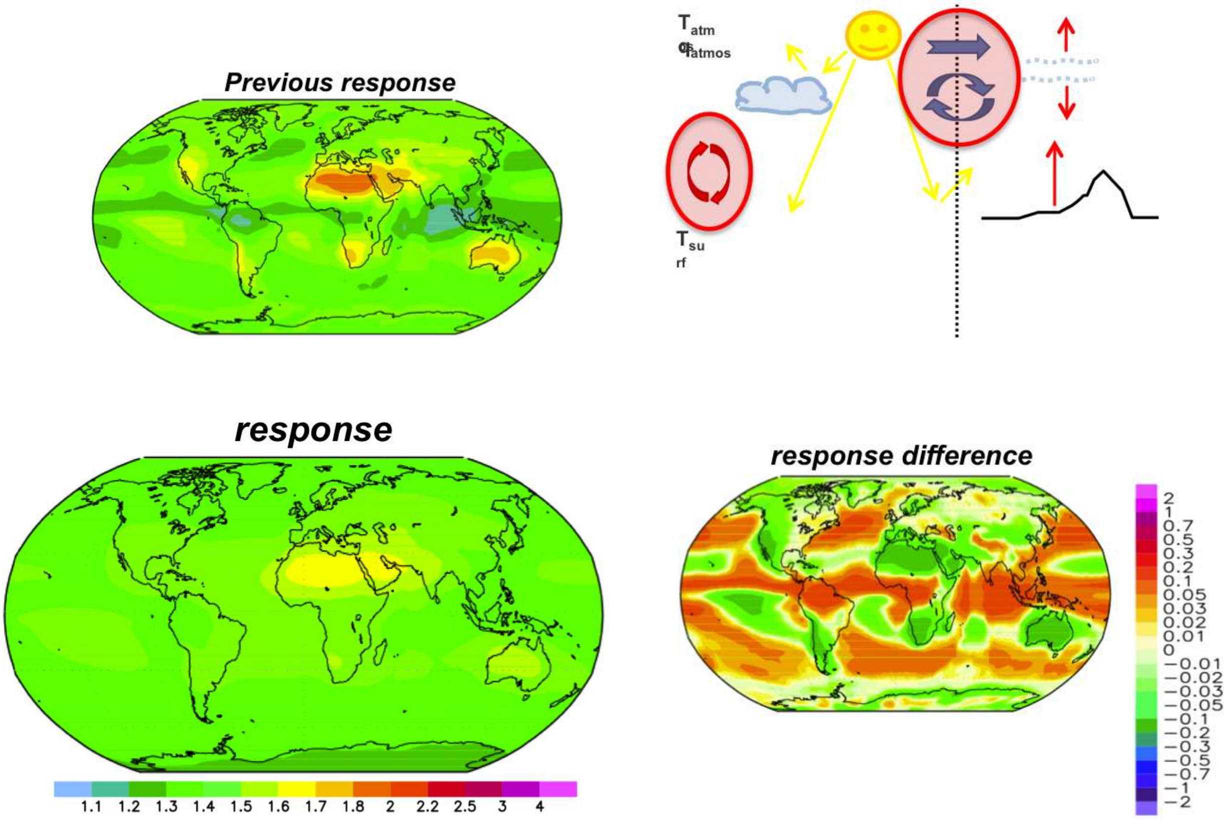

Figure 4.50: Exp. [5] Effect of atmospheric heat transport: No feedbacks. The response difference highlights the effect of the heat transport. Details as in Fig. 4.46.

- As [4] but with atmospheric circulation of heat
- No ocean heat uptake
  - $\rightarrow F_{ocean} = 0$
  - $\rightarrow T_{ocean} =$ fixed
- No feedbacks
  - $\rightarrow$ albedo = fixed
  - $\rightarrow q_{surf} =$ fixed
  - $\rightarrow$ No latent heating

The atmospheric heat transport by mean advection and diffusion reduces temperature gradients and extremes. The differences in the local sensitivities are smoothed out.

4.2.2 Ice/Snow-Albedo and Sea Ice Feedback

No water vapour feedbacks

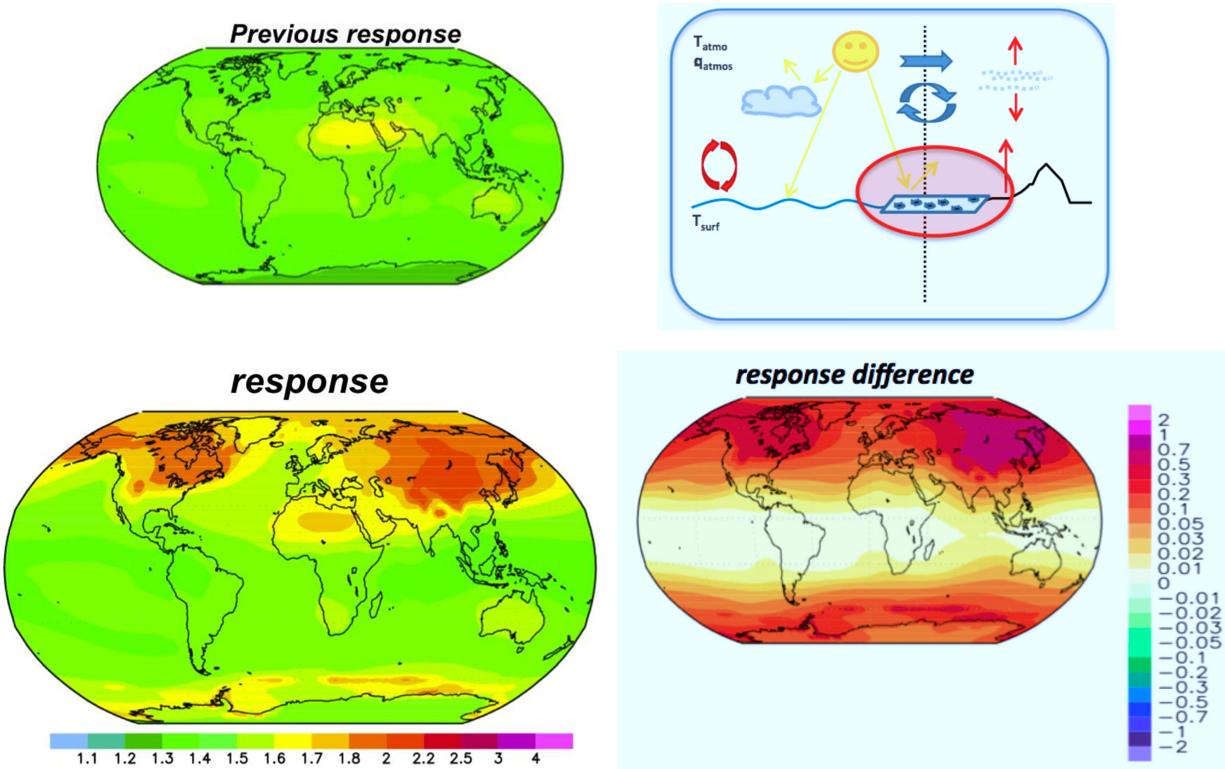

Figure 4.51: Exp. [6] Effect of ice/snow-albedo and sea ice feedback. No water vapour feedback. The response difference highlights the effect of the heat transport. Details as in Fig. 4.46.

- As [5] but with the ice-albedo and sea ice feedback.

- No ocean heat up take

In Sect. 4.1.2 we discussed the ice-albedo feedback. In Fig. 4.11 we illustrated that the ice-albedo feedback parameter is strongest in the midlatitudes of the northern hemisphere continents and around Australia. So the response to CO_2 forcing is amplified in these regions by the ice-albedo feedback to the warming T_{surf}.

4.2.3 The Water Vapour Feedback

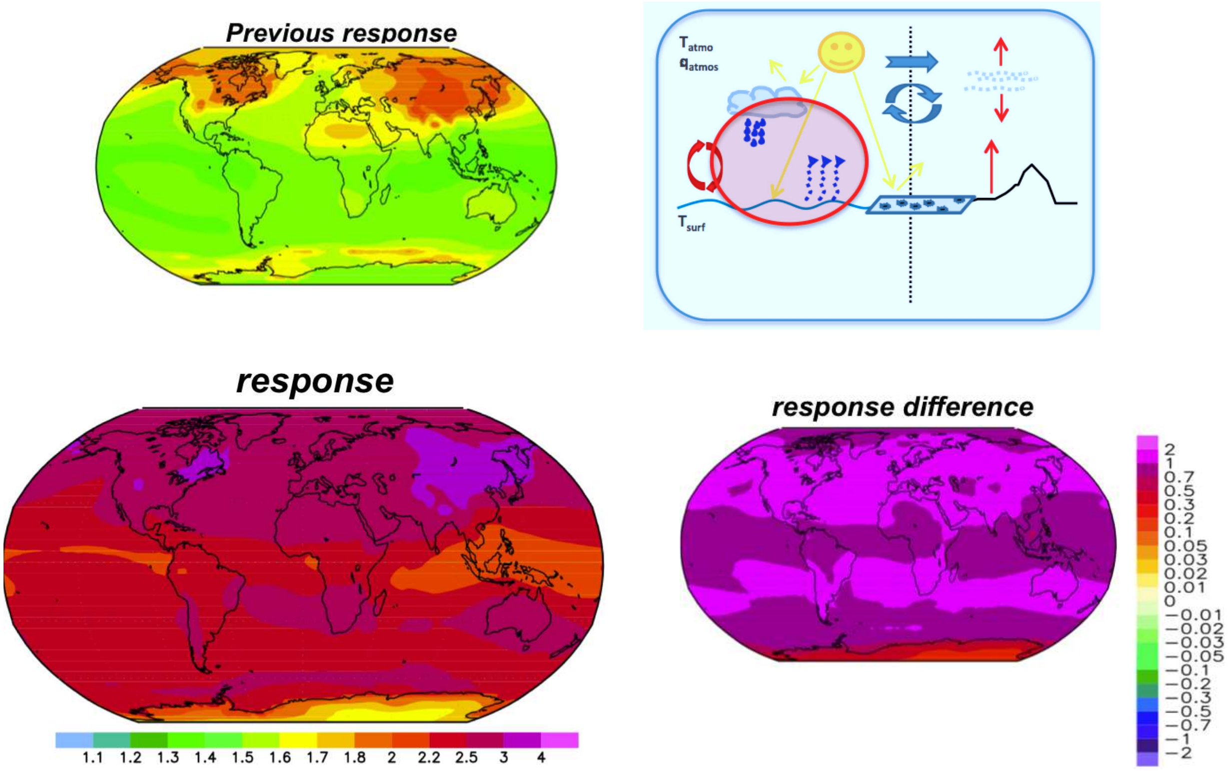

Figure 4.52: Exp. [7] Effect of local water vapour feedback: No transport of water vapour. The response difference highlights the local water vapour feedback. Details as in Fig. 4.46.

Experiment [7]: *Local Water Vapour Feedback*

- As [6] but with local water vapour response and latent heat release
- No ocean heat up take

The water vapour response to the warming is the most important feedback in the GREB model that amplifies the climate response to CO_2 forcing. The amount of water vapour in the atmosphere is strongly related to the temperature by the Clausius-Clapeyron equation (see eq. [4.14] and Fig. 4.22). A $3+^oC$ warming will increase the amount of water vapour in the atmosphere by 20% if the relative humidity is not changing. A 20% increase in the water vapour in the atmosphere changes the emissivity by about 0.02, see Fig. 4.53. This is as strong as the initial CO_2 effect and thus it basically doubles the climate response globally.

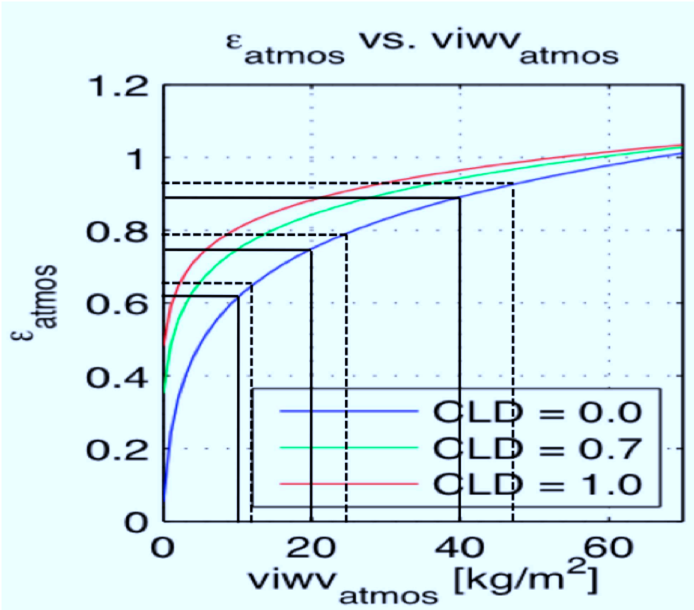

Figure 4.53: The sensitivity of the emissivity to changes in the total amount of water vapor in the atmosphere ($VIWV_{atmos}$). A 20% change in $VIWV_{atmos}$ will always lead to the same change in the emissivity.

Exp. [8] Effect of Turbulent Atmospheric Water Vapour Transport:

- As [7] but with turbulent (isotropic diffusion) water vapour transport
- No ocean heat up take
- Regions where water vapour increases a lot are now exporting additional water vapour to regions where water vapour increases less
- Warm and moist regions lose H_2O
- Cold and dry regions gain H_2O
- Dry and cold regions get additional warming by thermal radiation due to additional H_2O
- Warm and wet regions do the opposite
- There is also a minor effect of additional latent heating in the atmosphere by condensation of the additional H_2O

The local increase in the water vapour in the atmosphere is not uniform, warmer regions will have stronger increases in the water vapour. The atmospheric circulation will transport water vapour from regions with stronger local increase (warm and moist regions) to regions with smaller local increase (cold and dry).
The effect of the Turbulent Atmospheric Water Vapour Transport is therefore and additional warming in relatively dry and cold regions, see Fig. 4.54. These are in particular the land and polar regions. Thus the Turbulent Atmospheric Water Vapour Transport contributes to the land-sea warming contrast: the land warms more than the oceans.

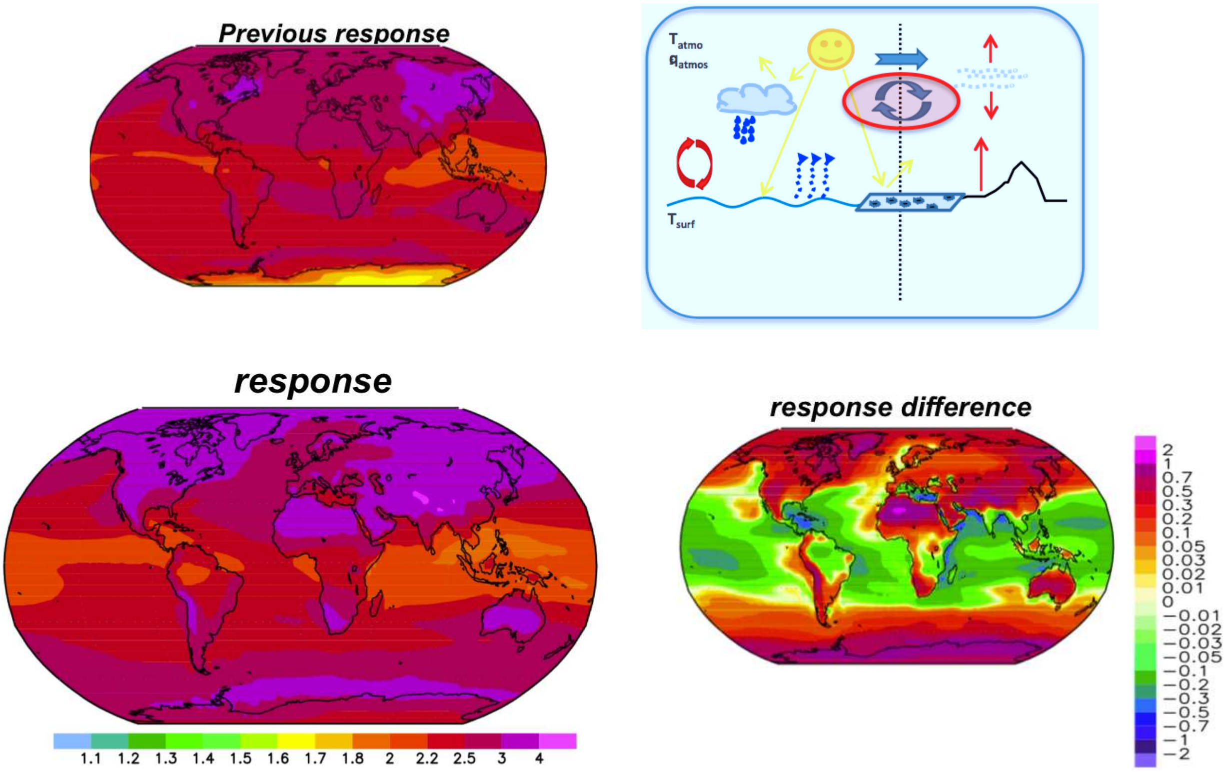

Figure 4.54: Exp. [8] Effect of turbulent atmospheric water vapour transport. The response difference highlights the effect of the heat transport. Details as in Fig. 4.46.

Exp. [9] Effect of Water Vapour Transport by the Mean Advection:

- As [8] but with mean advection transport of water vapour
- No ocean heat up take

The effect is similar to the turbulent atmospheric water vapour transport, but it is more focused on the northern hemisphere, because the mean winds blow more strongly across temperature and therefore water vapor gradients. In the southern hemisphere the more zonal wind do not blow across gradients in water vapor, see Fig. 4.55.

H

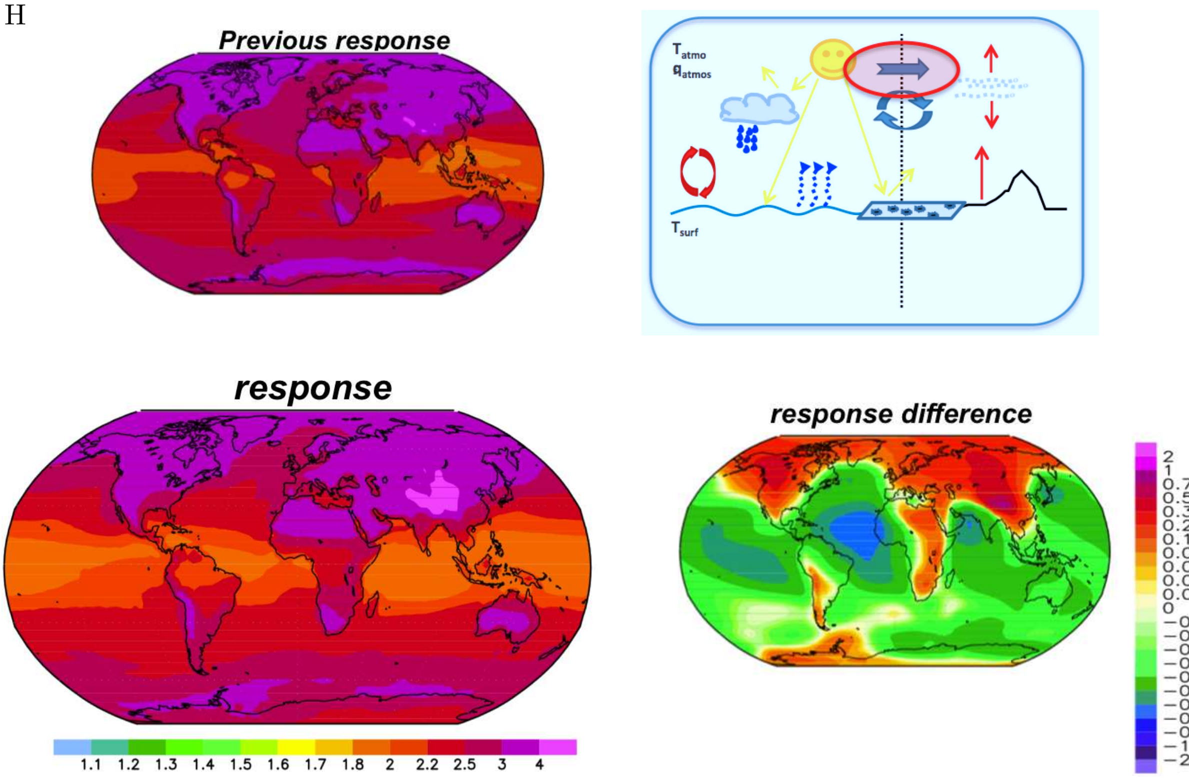

Figure 4.55: Exp. [9] Effect of Water Vapour Transport by the Mean Advection. The response difference highlights the effect of the heat transport. Details as in Fig. 4.46.

4.2.4 The Oceans Heat Up Take

Experiment [10]: Effect of deep ocean heat up take

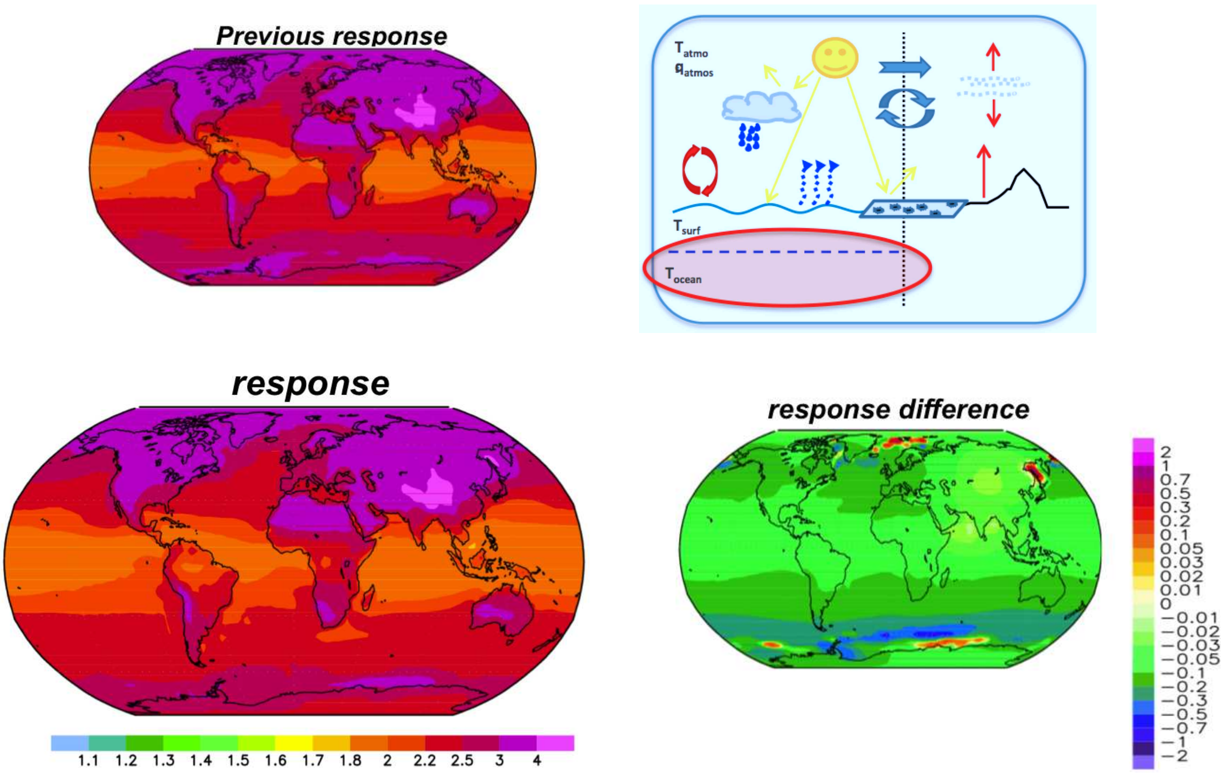

Figure 4.56: Exp. [10] Effect of the subsurface oceans heat up take. The response difference highlights the effect of the subsurface oceans heat up take. Details as in Fig. 4.46.

- As [9] but with deep ocean interaction
- The complete GREB model
- Ocean heat up take slows down the warming over oceans.
- The land feels the ocean response by atmospheric circulation of heat and moisture and therefore also slows down.
- How much the ocean slows down the warming depends on how fast the CO_2 concentration is changing.
- Ocean damping is a transient effect. In equilibrium the global mean is mainly not effected.
- The strongest effects are over the southern and northern North Atlantic Ocean, but in general the effect is global.
- The land vs. ocean warming ratio is a function of response time, but will always stay larger than 1.0, due to other feedbacks.
- the global mean resonse to 2xCO_2 forcing is 2.5 ° C after 50 years. In equilibrium it is 2.6 ° C, as in experiment [9].

- Note the regions of small increases in warming at the sea ice boundaries are due to changes of mixed layer depth in the ocean, which is an artifact of the GREB model experiments design.

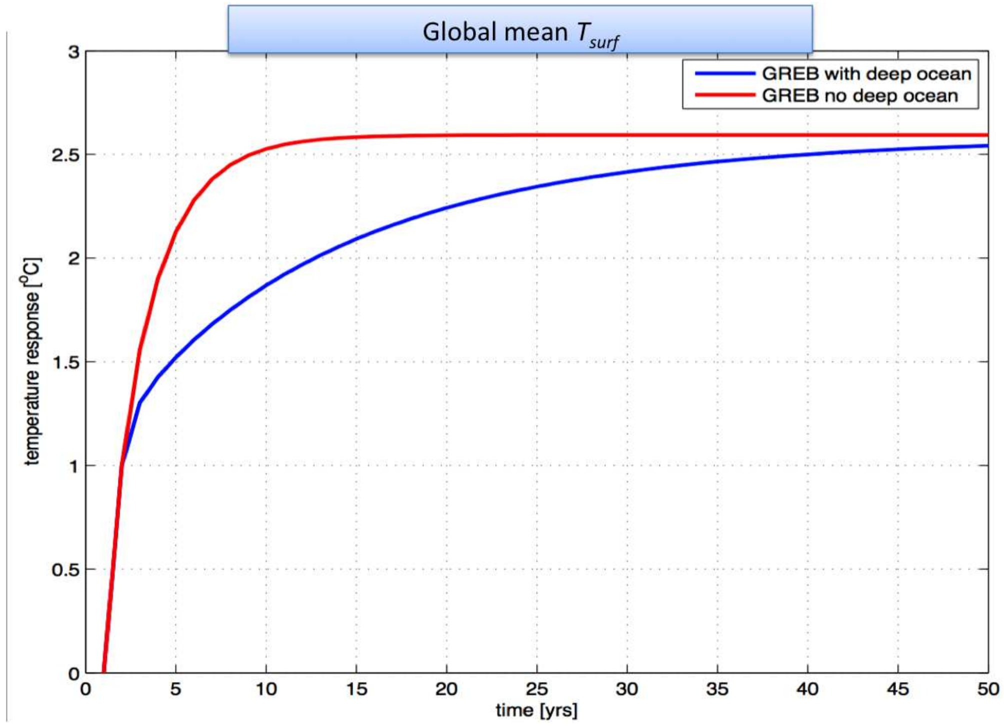

Figure 4.57: Effect of deep ocean heat up take - global mean surface temperature

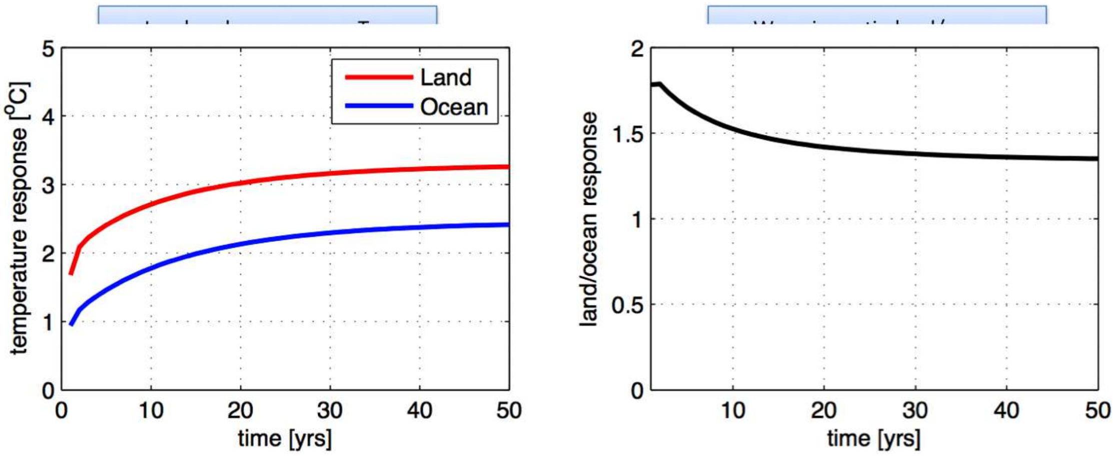

Figure 4.58: Effect of deep ocean heat up take - land vs. ocean

4.2.5 Cautionary Note on the GREB model global warming

- This is only a model, not the real world.
- The GREB is a very simple, uncertain and highly tuned (to IPCC models, not observed) model.

- IPCC model will in many regions or seasons have different responses for different reasons.
- The GREB model cannot say anything about how the circulation in the atmosphere or ocean responds. And it is likely that such responses will exist on the small scale (turbulence) and large scale (mean).
- GREB models cannot say anything about how cloud cover or soil moisture repond. Again this can be important.

4.3 IPCC-type climate models (General Circulation Models)

- IPCC-type climate models are based on weather forecast models
- Main features simulated by a weather/climate model:
  - Circulation of the atmosphere/ocean (dynamical core)
  - Radiation
  - Clouds
  - Land/Surface processes
  - Precipitation
  - Sea ice
- Each of these features is basically a model of their own
- As the core of these climate models is the dynamics of the circulation they are called ***General Circulation Models*** (GCMs). (GCM does not stand for: Global Climate Model!!)

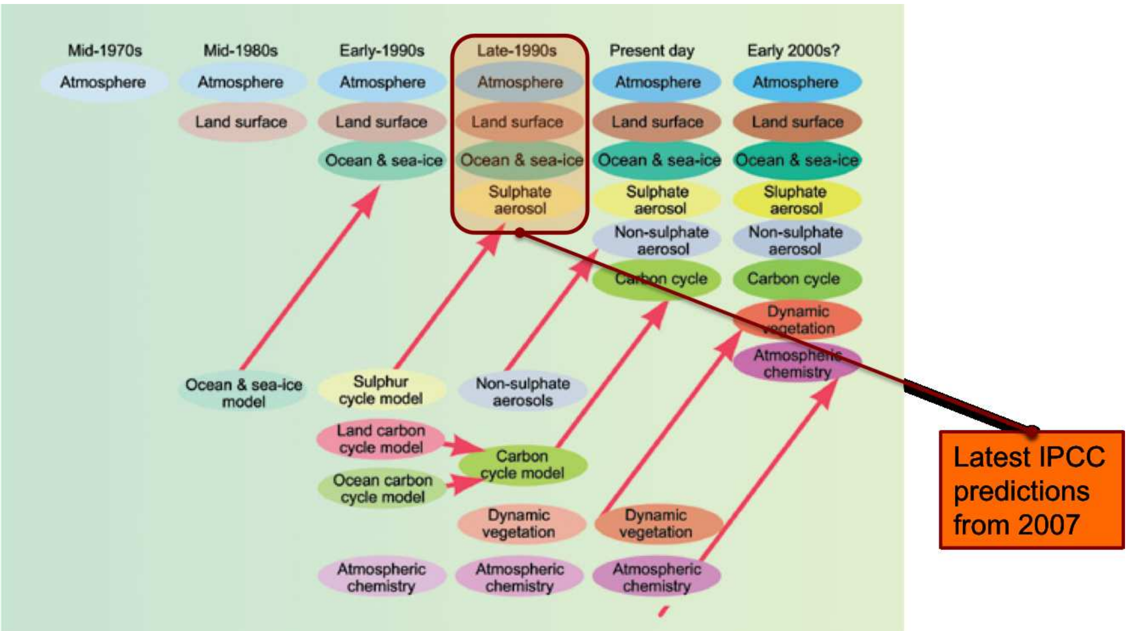

Figure 4.59: History of development of GCMs