

Environmental Physics Notes

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1 Introduction

1.1 Climate System Components

There are 5 main components of the climate system:

- **Atmosphere:** This has the fastest response time to changes in the climate system.
- **Lithosphere:** This is comprised of the crust and the upper mantle (the top layers of the Earth). It interacts mainly with the atmosphere through turbulent exchanges of mass e.g. desert dust storms, volcanic eruptions.
- **Hydrosphere:** This is comprised of all the water on Earth. It plays a key role on the climate by means of providing a heat reservoir due to its large mass and heat capacity. It also affects other components of the climate system through the water cycle and affects other cycles such as the carbon cycle.
- **Cryosphere:** This is comprised of all the ice on Earth (ice sheets, glaciers, snow fields, sea ice). It is important because it reflects a lot of solar radiation back into space. It also affects the albedo of the Earth (by reducing it if its cryosphere's size decreases and vice versa).
- **Biosphere:** This is comprised of all the living organisms on Earth. It affects the climate system by means of the carbon cycle and the water cycle. This is highly sensitive to climate state.

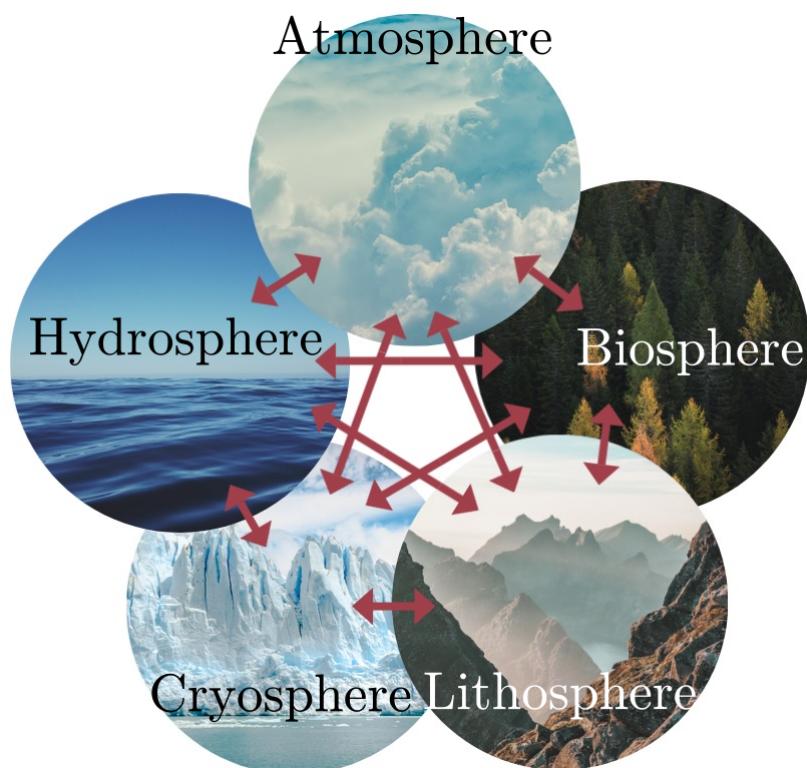


Figure 1: Climate System Components

1.2 Total Solar Irradiance

The climate system is fundamentally driven by the sun. Energy transfer from the Sun to the Earth is only done through EM radiation. To define this more rigorously, we assume the sun emmits **isotropically** (equally in all directions) as a **black body**. We can reach an expression for the irradiance I from the Sun reaching the TOA:

$$I = \sigma T_{\text{Sun}}^4 \frac{R_{\text{Sun}}^2}{R_{ES}^2} \quad \Rightarrow \quad \text{TSI} = \sigma T_{\text{Sun}}^4 \frac{R_{\text{Sun}}^2}{R_{ES}^2}$$

where σ is the Stefan-Boltzmann constant, T_{Sun} is the temperature of the Sun, R_{Sun} is the radius of the Sun and R_{ES} is the distance between the Earth and the Sun (radius of Earth's orbit, 1AU). A rough value for I is 1360 W m^{-2} .

Although the TSI is usually considered a constant, this does in fact vary on a number of timescales. There are millenial time-scale orbital variations (Milankovitch cycles) and the Sun's own activity cycle (11 years). The TSI can be measured in a number of ways, most recently by satellite measurements. The variation in the TSI seen over the last few decades is of the order 1-2 W m^{-2} .

1.3 Planetary Albedo and the Earth's Emitting Temperature

At the TOA, the only energy transfer is through EM (photon) radiation. Over time, the Earth will reach **radiative equilibrium** where the amount of energy absorbed by the Earth is equal to the amount of energy emitted by it at the TOA.

Not all the energy from the Sun is absorbed by the Earth. Some of it is reflected back into space. The fraction of energy reflected back into space is called the **planetary albedo**.

The Earth, with radius R_E and area πR_E^2 incident to the solar beam with albedo α_p will reach equilibrium when:

$$4\pi R_E^2 \sigma T_E^4 = (1 - \alpha_p)\pi R_E^2 \text{TSI} \quad \Rightarrow \quad T_E = \left(\frac{1 - \alpha_p}{4}\right)^{\frac{1}{4}} T_{\text{Sun}} \approx 255 \text{ K}$$

Here, we have equated the Earth's emission to the energy absorbed by it. Rearanging, we obtain the **emitting temperature** (T_E) of the Earth which comes out to about ~ 255 K. This is much lower than the actual average surface temperature of the Earth (~ 288 K). We will see later in the course that this is because the Earth in fact has an atmosphere which traps heat and increases the surface temperature as to balance the energy equation.

2 The Greenhouse Effect

Up to now we have considered a very crude model in which the Earth is a black body with no atmosphere, and its energy balance is ruled by the incoming solar radiation and the outgoing long wave radiation (LW). We have seen that such a model gave us a temperature that is significantly below the one that we observe. This is due to the **greenhouse effect**.

2.1 Emissivity, Absorptivity and Transmissivity

We can now come up with a more nuanced model of the Earth's energy balance. First we need to consider that materials are not perfect emitters, so we define the **Emissivity** ϵ_λ of a material as the ratio of the irradiance emitted by the material to the irradiance emitted by a black body at the same temperature. Notice that the Emissivity is a function of wavelength.

$$\epsilon_\lambda = \frac{I_{\text{material}}}{I_{\text{black body}}}$$

By definition, a black body has an Emissivity of $\epsilon_{BB} = 1$ so it follows that $I_{\text{material}} = \epsilon_\lambda \sigma T^4$. A material that emits like a **grey body** is one that has a constant Emissivity across all wavelengths.

We can also define the **Absorptivity** a_λ of a material as the ratio of the irradiance actually absorbed by the material to the irradiance incident on it. This is also a function of wavelength and is always a number between 0 and 1, meaning actual materials absorb less than a black body. It follows that, by definition, the Absorptivity of a black body is $a_{BB} = 1 = \epsilon_\lambda$ and in general (for systems in local thermodynamical equilibrium, as per Kirchoff's Law) $a_\lambda = \epsilon_\lambda$.

Finally, the more absorbing a material is, the less radiation it will transmit. We define the **Transmissivity** t_λ of a material as $1 - a_\lambda = 1 - \epsilon_\lambda$. This is also a function of wavelength and is always a number between 0 and 1. This relationship only holds in the absence of scattering, a reasonable assumption for clear-sky LW radiation.

2.2 Optical Depth

We can also write transmissivity in terms of the optical depth τ_λ of a material as follows:

$$t_\lambda = e^{-\tau_\lambda} \implies \tau_\lambda = \int_0^\infty k_{a\lambda} \rho_a dz$$

where $k_{a\lambda}$ is the **absorption coefficient** of the material, ρ_a is the density of the material across the atmosphere dz .

The optical depth is dimensionless: if the atmosphere has a τ of 1 it means that the amount of radiation transmitted through it from the surface to the TOA is reduced by a factor of $1/e$. Increasing the amount of absorbing material in the atmosphere will increase τ_λ and therefore reduce t_λ .

2.3 Definition of, and a Simple Model of the Greenhouse Effect

We will define the **greenhouse effect** (G) as the amount of LW radiation emitted by the Earth's surface that is trapped within the Earth's atmosphere. We can see how the atmosphere interacts with LW and SW radiation in figure 2.

As per figure 2, the atmosphere is much more transparent to SW radiation than it is to LW radiation. Water vapour is the main absorber of LW radiation followed by carbon dioxide and

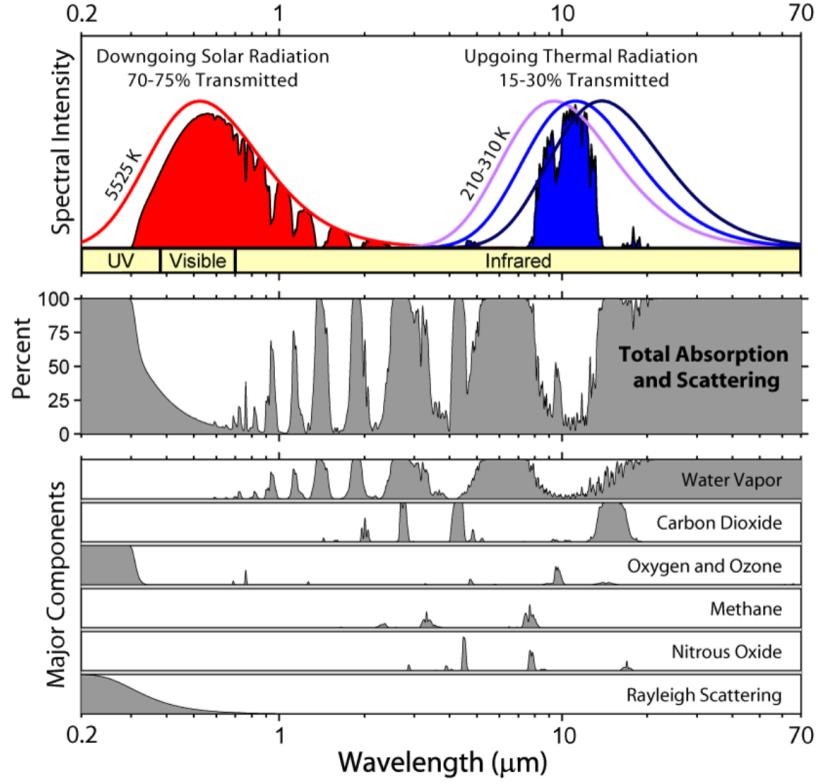


Figure 2: Radiation Transmitted by the Atmosphere.

some other gases. If we assume the Earth's surface has a temperature T_s and it emmits as a black body, then we can define G as:

$$G = \sigma T_s^4 - \text{OLR}$$

where OLR is the **outgoing long wave radiation** at the TOA, i.e. the sum total of radiation that escapes Earth into space.

Let us introduce a simple model of the greenhouse effect. We will have a few (important) assumptions:

- The atmosphere is completely transparent to **solar radiation (SW)**.
- We have blackbody emissioin from the surface at temperature T_s .
- The atmosphere is a single layer grey-body with Emissivity ϵ_a and temperature T_a emitting in the upward and downward directions equally.
- There is **radiative equilibrium** at the surface, in the atmosphere and at the TOA.

In this case, by energy conservation we can write the following for the **top of the atmosphere**:

$$\begin{aligned} \text{Incoming SW from the Sun} &= \text{LW emitted by the atmosphere} + \text{Transmitted LW from the surface} \\ \frac{(1 - \alpha_p) \text{TSI}}{4} &= \epsilon_a \sigma T_a^4 + (1 - \epsilon_a) \sigma T_s^4 \end{aligned}$$

within the atmosphere we can write:

$$\begin{array}{lcl} \text{LW emitted by the} \\ \text{surface absorbed by atmos} \\ \epsilon_a \sigma T_s^4 \end{array} = \begin{array}{l} \text{LW emitted by the} \\ \text{atmosphere up and down} \\ 2\epsilon_a \sigma T_a^4 \end{array}$$

finally, **at the surface** we can write:

$$\begin{array}{lcl} \text{LW emitted} \\ \text{by the surface} \\ \sigma T_s^4 \end{array} = \begin{array}{l} \text{Incoming} \\ \text{SW from the Sun} \\ \frac{(1 - \alpha_p) \text{ TSI}}{4} \end{array} + \begin{array}{l} \text{LW emitted by the} \\ \text{atmosphere downwards} \\ \epsilon_a \sigma T_a^4 \end{array}$$

With these we have a set of simultaneous equations with which we can solve for T_s and T_a for some given reasonable values for α_p , ϵ_a and TSI. We can solve for T_s and G as follows:

$$\begin{aligned} T_s &= \left(\frac{(1 - \alpha_p) \text{ TSI}}{2\sigma(2 - \epsilon_a)} \right)^{1/4} \\ \text{OLR} &= \epsilon_a \sigma T_a^4 + (1 - \epsilon_a) \sigma T_s^4 \\ G &= \sigma T_s^4 - \text{OLR} \\ &= \epsilon_a \sigma T_s^4 - \epsilon_a \sigma T_a^4 \end{aligned}$$

having used the definition of $G = \sigma T_s^4 - \text{OLR}$ and the first equation above for the TOA balance being in equilibrium (i.e. OLR = incoming SW from the Sun).

With a further substitution from the atmospheric balance (2nd equation above) for T_a we can write:

$$G = \frac{\epsilon_a}{2} \sigma T_s^4$$

which is what we expect: a thicker, more absorbing atmosphere will have a lower Transmissivity meaning a higher Emissivity (ϵ_a) meaning a stronger greenhouse effect G .

3 Radiative Forcing and Feedback

3.1 Radiative Forcings and Feedbacks

Let I_N be the net irradiance at the TOA, defined positive in the downward direction. It follows that in **radiative equilibrium**:

$$I_N = \frac{(1 - \alpha_p)TSI}{4} - OLR = 0$$

- A **radiative forcing** (ΔQ_{EXT}) is an external perturbation to the climate system away from radiative equilibrium at the TOA. A positive forcing leads to a positive global mean surface temperature change (ΔT_s), and vice versa.
- A **feedback** (ΔQ_{INT}) is an internal process which responds to the forcing **and** has an effect on the global mean surface temperature.

It follows from the above definitions that:

$$\text{In the absence of feedback processes } \Delta I_N = \Delta Q_{EXT}$$

$$\text{In the presence of feedback processes } \Delta I_N = \Delta Q_{EXT} + \Delta Q_{INT}$$

We can also define a further parameter, the **climate feedback parameter** λ . This is a measure of how much the planet's energy balance responds to changes in global mean surface temperature ΔT_s . It is a measure of how much additional energy is trapped or lost as a result to said change in surface temperature. We define the total **climate feedback parameter** γ as:

$$\Delta Q_{INT} = \gamma \Delta T_s$$

where γ represents the radiative response of the climate feedback to a given ΔT_s . Its units are $\text{W m}^{-2}\text{K}^{-1}$.

3.2 Decomposing the Feedback Parameter

We can decompose the total feedback parameter γ into the sum of individual feedback parameters:

$$\gamma = \frac{dI_N}{dT_s} = \frac{\partial I_N}{\partial T_s} + \sum_x \frac{\partial I_N}{\partial x} \frac{\partial x}{\partial T_s}$$

where x represents the various parameters (e.g. water vapour). Here, $\frac{\partial I_N}{\partial T_s} = -\gamma_{BB}$ is the **blackbody feedback parameter**. The negative sign is simply to represent the fact that the surface will radiate more energy as it warms, reducing the net irradiance at the TOA (I_N).

Example:

For a simple 1 layer atmosphere transparent to solar radiation, it follows that $OLR = \epsilon' \sigma T_s^4$ where ϵ' is the effective emissivity which reduces with atmospheric absorptivity and hence emissivity increases. We can write:

$$\frac{\partial I_N}{\partial T_s} = -\frac{\partial OLR}{\partial T_s} = -\gamma_{BB} \implies \boxed{\gamma_{BB} = 4\epsilon' \sigma T_s^3}$$

As per the example above, it becomes apparent that for a parameter (x) to exert a strong feedback, it must be dependent on both T_s **and have an influence on** the net irradiance at the TOA (I_N).

3.3 Water Vapour Feedback and Clausius-Clapeyron Scaling

We mention in section 2.3 that water vapour is the main absorber of LW radiation, and therefore the most important greenhouse gas. Water vapour concentrations show a strong dependence on temperature, i.e. the warmer the atmosphere, the more water vapour it can hold. This is a result of the Clausius-Clapeyron relation.

As per Paterson's thermodynamics course, we know that the pressure of water vapour which is saturated with respect to liquid water at a given temperature T is given by:

$$\frac{dP_s}{dT} = \frac{s_v - s_l}{v_v - v_l} \implies \boxed{\frac{dP_s}{dT} = \frac{l_v}{T v_v}}$$

at the vapour-liquid phase boundary where s_v and s_l are the specific entropies of vapour and liquid water respectively, v_v and v_l are the specific volumes of vapour and liquid water respectively, l_v is the latent heat of vaporisation and P_s is the saturation vapour pressure.

If we assume water vapour behaves as an ideal gas, it is apparent that $P_s v_v = N k_B T / N m_v = k_B T / m_v$ where m_v is the molecular mass of water (remember we are using specific volume, not volume). We can define the gas constant for water vapour $R_v = R_v = k_B / m_v$ and hence:

$$\frac{dP_s}{dT} = \frac{P_s l_v}{R_v T^2} \implies \frac{dP_s}{P} = \frac{l_v}{R_v T^2} dT$$

for context, we find that for sensible values for the constants:

$$\left(\frac{\frac{dP_s}{P_s}}{dT_s} \right) \approx 0.07 \text{ K}^{-1}$$

we call this the Clausius-Clapeyron scaling. Note that this does not apply to the actual vapour pressure (related to the absolute amount of water vapour in the atmosphere). Vapour pressure and saturation vapour pressure are related by:

$$\text{Relative Humidity, RH} = \frac{P_v}{P_s} \times 100\% \implies +7\% \text{ RH per degree warming}$$

where P_v is the vapour pressure.

Climate models suggest that relative humidity is conserved under anthropogenic climate change meaning that the absolute amount of water vapour in the atmosphere increases as atmospheric temperature increases.

4 Water in the Climate System

Figure 3 shows us the radiative feedback of various parameters (see section 3). In this section we will focus on how these parameters lead to a positive or negative radiative feedback.

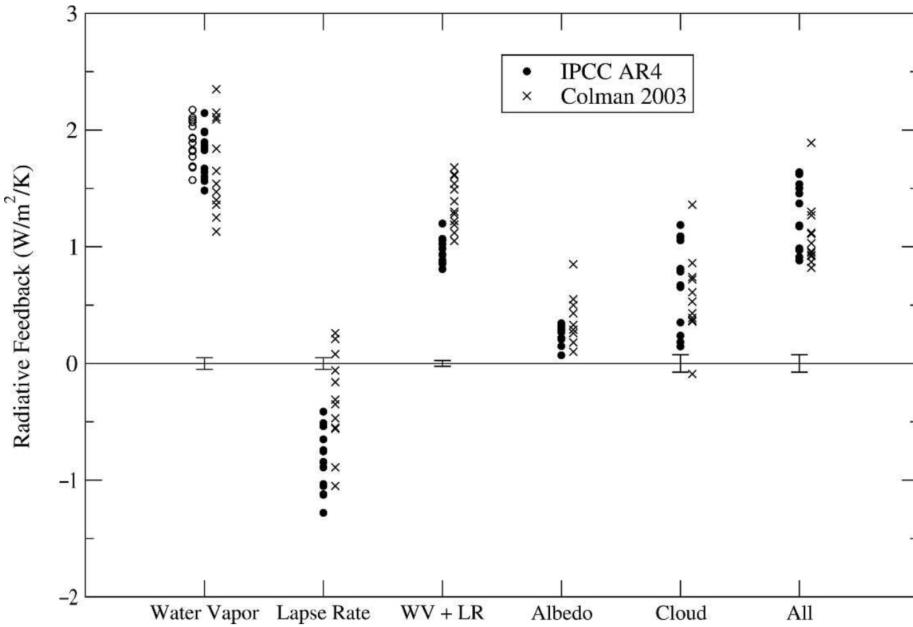


Figure 3: Radiative Feedback of Various Parameters ([1, Soden and Held 2006]).

4.1 Ice-Albedo Feedback

The distribution of sea-ice shows a natural seasonability linked to the pattern of solar radiation at the surface with a maximum extent at both poles after their respective polar nights (and vice versa). Over the last few decades, satellite observations have shown a clear decrease in the extent of sea-ice in the arctic (this is not the case in the Antarctic, where the pattern seen is different). There is no clear sign of decreasing of sea-ice, in fact there are signs of a slight increase. The reasons why are still up for debate and more recent data post 2017 show a decrease as expected with the warming of the climate. This is non-examinable).

The **ice-albedo feedback** mechanism suggests that an increase in surface temperature would result in a reduced snow and sea-ice cover and hence reduce surface albedo (α). This would result in a decrease in planetary albedo α_p and hence an increase in absorbed solar radiation. This would lead to a further increase in surface temperature and so on. This is a **positive feedback** mechanism as shown in Figure 3.

4.2 Lapse Rate

The **lapse rate** is the negative rate at which temperature changes with height in the atmosphere. In the troposphere, where the temperature decreases with height, the lapse rate is positive. We define lapse rate Γ as:

$$\Gamma = -\frac{dT}{dz}$$

where T is temperature and z is height.

For a parcel of dry air lifted in an atmosphere due to convection, it follows that it does so adiabatically. In this case we say that it rises at the **dry adiabatic lapse rate** Γ_d .

$$c_P dT - v dP = dq \quad \text{1st Law of Thermodynamics}$$

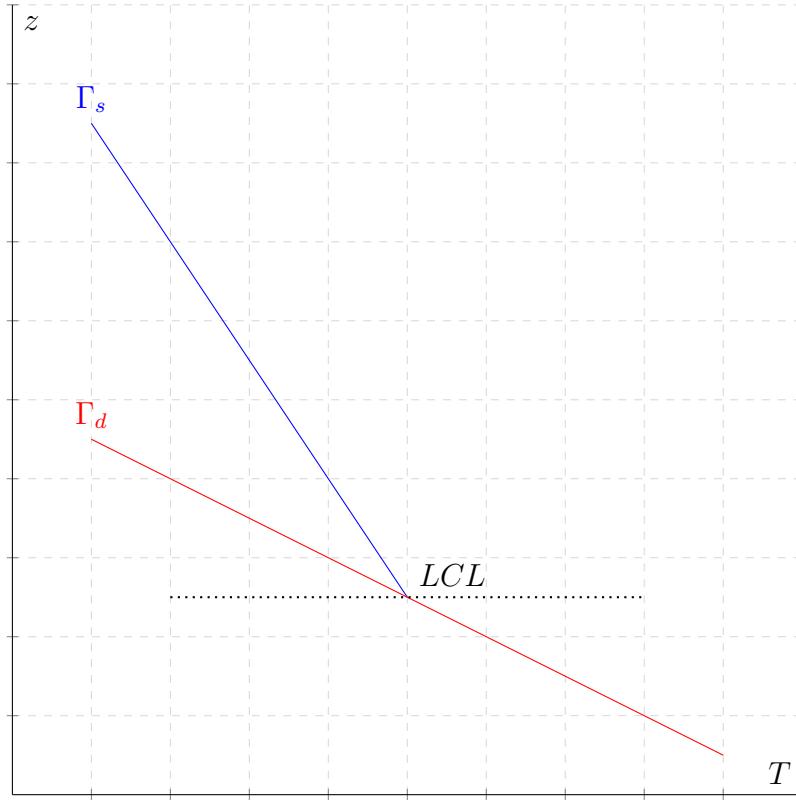


Figure 4: Dry and Moist Adiabatic Lapse Rates.

$$\frac{dT}{dz} = -\frac{g}{c_P} \quad \Rightarrow \quad \boxed{\Gamma_d = \frac{g}{c_P}}$$

The same thing happens for a parcel of moist air. We begin by imagining a parcel of air near the surface that contains some non-zero amount of water vapour. As the parcel rises, it cools adiabatically so it gets closer and closer to its saturation point. Eventually it does, and the height at which it reaches the saturation point is called the **lifting condensation level (LCL)**. At this point, the parcel condenses and releases latent heat as it rises further. This heat means the parcel's temperature does not decrease as fast as it otherwise would. Such a parcel is said to be following the **moist adiabatic lapse rate Γ_s** .

We can combine the two lapse rates into one equation as follows (see handout 2, moist adiabatic lapse rate for derivation):

$$\Gamma_s = \Gamma_d \frac{1 + \frac{w_s l_v}{R_d T}}{1 + \frac{\epsilon l_v^2 w_s}{c_p R_d T^2}} = \Gamma_d \frac{1 + X}{1 + bX}$$

where l_v is the latent heat of vapourisation, w_s is the **saturation mixing ration**, T is the temperature, R_d is the gas constant for dry air, c_p is the specific heat capacity of dry air at constant pressure, ϵ is the ratio of the gas constants for dry air and water vapour, $X = \frac{w_s l_v}{R_d T}$ and $b = \frac{\epsilon l_v}{c_p T}$.

To summarise, and worth noting is the following:

$$\boxed{\Gamma_d = \frac{g}{c_p}, \quad b > 1, \quad \Gamma_s < \Gamma_d}$$

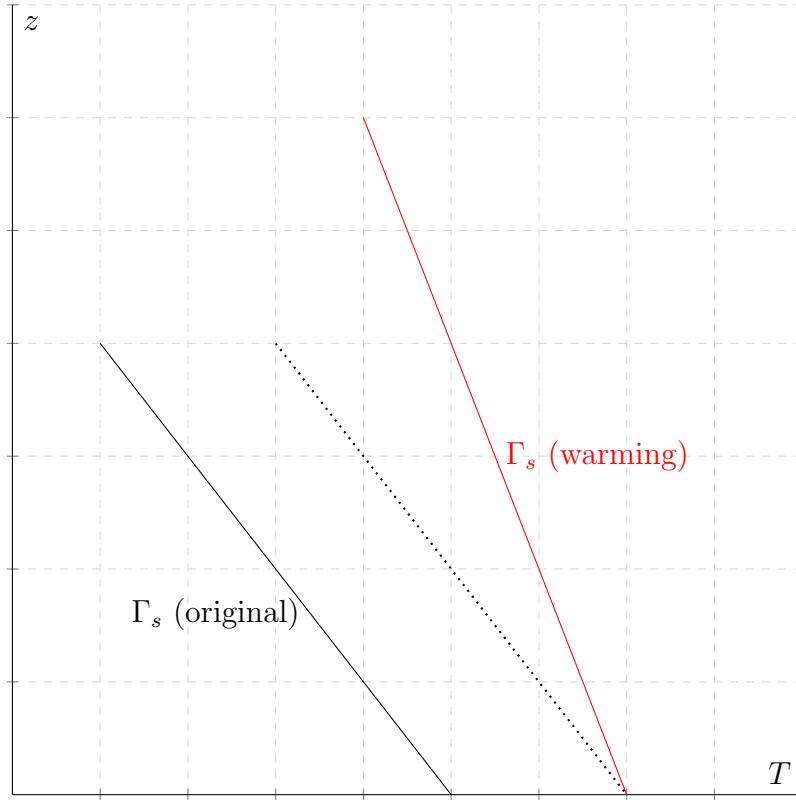


Figure 5: Lapse Rate Feedback.

4.3 Lapse Rate Feedback

We showed in section 4.2 that the rate of change of Γ_s with temperature must be negative. In regions of high moisture and convective uplift (e.g. tropical troposphere) a change in surface temperature will in turn lead to a change in the atmospheric temperature at higher altitudes. If we have a radiative forcing which results in a surface warming, this will lead to a decrease in Γ_s (the graph in Figure 5 gets steeper). In other words the temperature at higher altitudes will increase more than “expected”, the graph does not simply translate. We can see this in Figure 5.

Figure 5 shows that the lapse rate feedback is negative. This means that a radiative forcing will lead to a greater change in temperature at higher altitudes than expected. This is a positive feedback because the change in temperature at higher altitudes will lead to a greater LW emission at the TOA (equivalently, an increase in OLR) and also a net decrease in LW emission back towards the surface. This will therefore lead to a cooling of the surface, hence negative feedback.

4.4 Lapse Rate and Water Vapour Feedback Combined

We have seen in section 3.3 that relative humidity tends to be conserved i.e. if atmospheric temperature increases or decreases, the relative humidity stays the same meaning the absolute levels of water in the air increase or decrease respectively. This is a consequence of Clausius-Clapeyron scaling (again, see section 3.3 for the details).

A higher absolute level of water vapour in the air also means a higher latent heating once saturation point is reached (there is more water per unit volume that condenses, so more heat released per unit volume). This means that the water vapour and lapse rate are intimately related. More water vapour means a lower lapse rate due to increase in temperature, and a lower lapse rate means more heat is radiated out to space. Overall the combined feedback is

positive, see figure 3.

4.5 The Impact of Clouds on Radiation

Clouds interact strongly with both SW and LW radiation. They are highly reflective to SW radiation (they are responsible for 50% of the Earth's Planetary Albedo) and are also strong absorbers and emitters of LW. However the net impact is highly dependent on a few factors.

In the presence of sunlight, clouds have a very reflective effect on SW radiation compared to clear-sky conditions, especially if they are over a dark surface (e.g. ocean or vegetation).

If the cloud is at a low altitude, this cloud is hot and therefore a good blackbody emitter so there is a net loss of energy to space. The cloud therefore cools the surface.

If the cloud is at a high altitude, this cloud is cold and therefore does not emit much LW radiation. The LW effect will dominate over the SW effect and the cloud will act as a blanket, trapping heat and warming the surface.

At night, all clouds (due to their relatively high optical thickness) they always act as a blanket, trapping heat and warming the surface (or rather not letting it cool as much as it otherwise would).

In general we will think of a low cloud = cooling, high cloud = warming with the exception of night time we noted before.

Note that the above is a very simplified picture. In reality, clouds are highly variable in their optical thickness (our main assumption was that clouds are very thick) and they have all sorts of microphysical properties which affect their feedback properties. The value of the cloud feedback parameter γ_c is the most uncertain one and varies significantly between models.

From our simple model we can write the following equation for the cloud feedback parameter:

$$\gamma_c = \frac{\partial I_N}{\partial x} \frac{\partial x}{\partial T_s}$$

where the first term depends on cloud altitude, time of day/year, cloud microphysics and otherwise. The second term is a non-trivial relationship between cloud properties and temperature.

5 The Carbon Cycle

5.1 Fast and Slow Carbon Cycle

- **Fast Carbon Cycle:** This is the flow of carbon from a reservoir or store to another where each cycle takes about 10s of years. This tends to include the atmosphere, biosphere, soils and upper ocean. These stores are relatively small.

Anthropogenic changes in the stores are relatively small when compared to natural cycles, but still disturb the natural balance of the cycle

- **Slow Carbon Cycle:** This is the flow of carbon from a reservoir or store to another where each cycle takes 100,000s of years. This tends to include the deep ocean and geological sediments (e.g. fossil fuel reservoirs or carbonate rocks like limestone). These stores are relatively large.

5.2 Atmospheric Carbon Dioxide Concentrations, Emissions and Up-take

CO₂ concentration is something that has changed over time due to natural and anthropogenic factors. Concentrations are higher today than they have been at any point in the last 400,000 years. Over the last 1000 years up until 1750 (the start of the industrial revolution) CO₂ concentrations were relatively stable at ~285ppm. Since then they have increased to the current level of ~415ppm.

Over this period, CO₂ has been the largest contributor to the total anthropogenic radiative forcing. This is a result of the increase in CO₂ concentration in the atmosphere, its strong absorption peak at 15μm and its long atmospheric lifetime.

Anthropogenic emissions of CO₂ are mainly due to the burning of fossil fuels and cement production. We can combine the net increase in CO₂ concentration in the atmosphere by considering the emissions and uptake of CO₂ in the three main sinks: the atmosphere, the ocean and the land biosphere. Furthermore, from this we can also consider the **airborne fraction** of CO₂ emissions (the fraction of emissions that remain in the atmosphere).

Research suggests that the land biosphere and ocean's ability to absorb CO₂ has increased over time, more than doubling since 1960. Without this, atmospheric concentrations of CO₂ would be even higher than they are today.

Over the past decade, of the ~40 GtCO₂ emitted each year, 46% has remained in the atmosphere, 31% has been absorbed by the land biosphere and 23% has been absorbed by the ocean.

5.3 The Terrestrial Carbon Cycle

5.3.1 Photosynthesis and Respiration

Photosynthesis and respiration are the two main processes that control the exchange of CO₂ between the atmosphere and the land biosphere. Photosynthesis *removes* CO₂ from the atmosphere and is energy intensive and respiration *adds* CO₂ to it releasing energy.

Plants do not absorb all light equally when photosynthesising (hence them not being black). They are most sensitive between 400-700nm. The absorption of light is used to drive the **Photosynthetically Active Radiation** (PAR): this is the integral of the graph seen in figure 6. In said figure we can see a dip around the 550nm region, which is green colour light, hence why

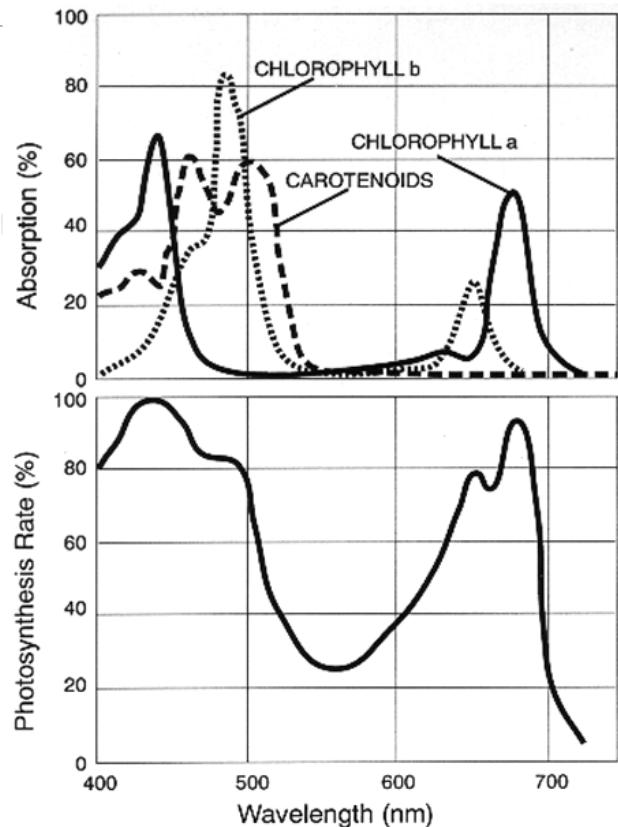


Figure 6: The PAR Action Spectrum

plants are green.

We can also define the productivity of an ecosystem with the **Gross Primary Productivity** (GPP) and the **Net Primary Productivity** (NPP). GPP is the total amount of carbon fixed by photosynthesis in an ecosystem per unit time, whereas NPP is the GPP minus the amount of carbon released by respiration i.e. $NPP = GPP - R$.

Around 50% of the GPP is used for respiration, the other 50% is used for growth.

5.3.2 The Terrestrial Biosphere as a Carbon Source and Sink

As noted in section 5, the terrestrial biosphere is a sink for CO_2 , both from natural and anthropogenic origin. Natural variability of the terrestrial biosphere occurs and can perturb the magnitude of this sink.

However, anthropogenic activity has a significant impact. Indeed, land use and change through deforestation and some agricultural practices have resulted in the terrestrial biosphere becoming a net source of CO_2 in some regions (e.g. the tropics). On the contrary, in other regions this is in fact the opposite, where afforestation and reforestation have resulted in the terrestrial biosphere becoming a net sink of CO_2 and some regions even have net negative land carbon emissions.

It is important the permanence of the carbon sink is considered when thinking about the influence of the terrestrial biosphere on atmospheric CO_2 concentrations. For example, wildfires can perturb concentrations, but vegetation tends to regrow and reabsorb the CO_2 released, meaning this is a transitory effect.

5.3.3 Climate-Carbon Feedbacks

Both photosynthesis and respiration are temperature dependent. As such changes to the climate will perturb the terrestrial carbon cycle and therefore the atmospheric CO₂ concentration. Some of these feedbacks follow:

- ‘Greening’ of the Earth due to the **CO₂ Fertilisation Effect**: This is the increase in photosynthesis due to the increase in CO₂ concentration in the atmosphere. This is a negative feedback as the increase in photosynthesis will result in a decrease in atmospheric CO₂ concentration, which will in turn result in a decrease in photosynthesis. This enhances the land carbon sink.
- Relaxation on factors limiting plant growth - water, sunlight availability and temperature.
- Increase in respiration due to the increase in temperature. This is a positive feedback as the increase in respiration will result in an increase in atmospheric CO₂ concentration, which will in turn result in an increase in temperature which leads to even more respiration. This limits the land carbon sink.

The impact of these is non-trivial and is still an area of active research.

5.4 The Oceanic Carbon Cycle

The ocean is a large carbon sink for atmospheric CO₂. It takes up CO₂ through the dissolution of CO₂ in the surface ocean and transforms it into **Dissolved Inorganic Carbon** (DIC). We can define DIC as:

$$\begin{aligned} \text{DIC} &= \text{Dissolved Carbon Dioxide} + \text{Carbonic Acid} + \text{Bicarbonate} + \text{Carbonate} \\ \text{DIC} &= \text{CO}_2 \text{ (aq)} + \text{H}_2\text{CO}_3 + \text{HCO}_3^- + \text{CO}_3^{2-} \end{aligned}$$

As we can expect, DIC has increased over the last few decades due to the increase in anthropogenic CO₂ emissions. As per the equation above we can see that this increase in DIC leads to a decrease in pH, i.e. ocean acidification. This is partially buffered but oceanic acidification is still a concern and therefore a concern.

5.4.1 Controls on Oceanic Carbon Uptake

How much the oceans are able to absorb CO₂ is dependent on two main factors: atmospheric CO₂ concentration and oceanic temperature:

- Oceans are able to absorb more CO₂ as DIC when atmospheric CO₂ concentration is higher. However, the relationship between atmospheric CO₂ concentration and DIC is not linear, meaning the oceans' ability to keep absorbing CO₂ as concentrations increase more and more is reducing.
- Oceans are able to absorb more CO₂ as DIC when oceanic temperature is lower. This is because the solubility of CO₂ in water is higher at lower temperatures. It therefore follows that increasing oceanic temperature reduces the oceans' ability to act as carbon sinks.

5.4.2 The Ocean Biological Carbon Pump

This is the process by which carbon is transported from the surface ocean to the deep ocean. It is a biological process where the transformation of DIC into organic carbon occurs through photosynthesis of marine plants and phytoplankton. These living organisms then die and sink to the deep ocean. During this process, some of the organic carbon will decompose and be released back into the ocean as DIC. However, some of it will be buried in the deep ocean, which is a carbon sink.

The remineralisation depth, or the depth at which 63% of the organic carbon has been remineralised determines the time interval until the carbon is released back into the atmosphere. A shallower depth will reduce the capacity of the ocean to draw down anthropogenic carbon emissions.

6 Other Climate Forcers

6.1 Well-Mixed Greenhouse Gases

We have seen in previous sections that carbon dioxide is the main forcing agent of anthropogenic climate change, however, there are several other important players. These are a variety of gases that are grouped into what is called the **well-mixed greenhouse gases** (WMGHG). These are gases that are well-mixed in the atmosphere (troposphere), i.e. their concentrations do not vary significantly with geographical location or height. The main WMGHG are methane, CFC-11 and CFC-12.

Methane is the second most important anthropogenic greenhouse gas. It is emitted from a variety of sources, both natural and anthropogenic. Its atmospheric lifetime is ~ 12 years and it is removed from the atmosphere by reaction with the OH radical. Its concentration rate increase is of $\sim 18\text{Tg CH}_4\text{yr}^{-1}$, i.e. emissions increase by $\sim 10\%$ per decade (has been for the past two decades).

CFC-11 and CFC-12 are **chlorofluorocarbons** (CFCs) and are synthetic gases, part of the family of Halocarbons (hydrocarbons that have had some or all of their hydrogen atoms replaced by halogen atoms). They are emitted from anthropogenic sources and have lifetimes of ~ 50 and ~ 100 years for CFC-11 and CFC-12 respectively.

These are inert in the troposphere but dissociate in the stratosphere leading to ozone depletion. Environmental and health concerns have led to the Montreal Protocol in 1987 ultimately prohibited CFC production by 2000 leading to a decrease in their atmospheric concentrations.

6.2 Aerosol Radiative Forcing

Aerosols are small liquid drops or particulates suspended in the atmosphere. They can be natural or anthropogenic in origin and can be made up of many different constituents. They can be emitted directly into the atmosphere (primary) or formed in the atmosphere as a result of chemical processes (secondary).

Aerosols have a direct effect on the climate by scattering and absorbing radiation. They also have an indirect effect by modifying the properties of clouds.

Some aerosols like black carbon or soot have a predominantly warming effect since they absorb radiation thus causing them to warm up. More commonly, aerosols have a predominantly cooling effect since they scatter radiation back to space, examples of this are sulphate aerosols and sea salt aerosols. It is for this reason that aerosols are believed to have had a cooling effect on the climate for the past 250 years or so.

Some aerosols like sulphate aerosols can also act as **cloud condensation nuclei** (CCN) and thus modify the properties of clouds. If more aerosols are present, then more CCN are present and thus more cloud droplets are formed (smaller droplets) and thus the cloud is brighter and more reflective. This is known as the **first aerosol indirect effect** or **Twomey effect**.

Other indirect effects link aerosol to changes in precipitation. The smaller droplets seen above take longer to grow into raindrops, delaying precipitation and prolonging the cloud lifetime. Both of these effects are believed to have a cooling effect on the climate.

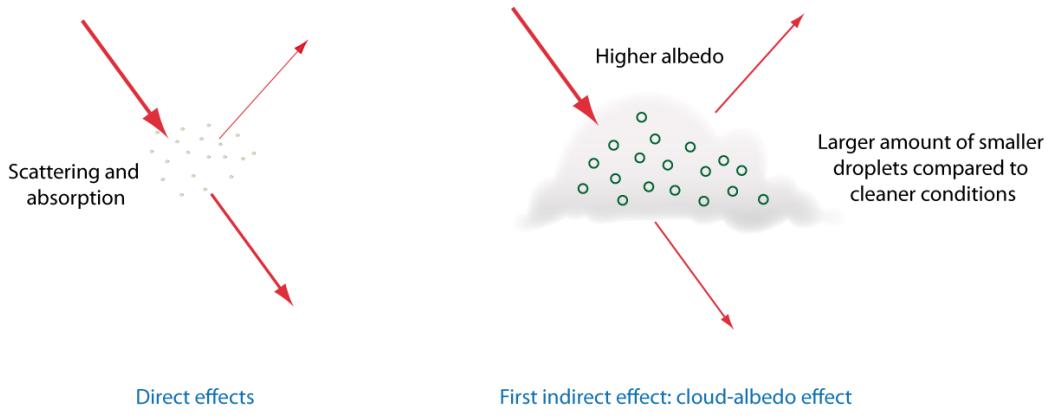


Figure 7: Direct and indirect effects of aerosols on the climate.

6.3 Comparing the Impact of Different Forcing Agents

To determine the impact a forcing agent (radiative gas or aerosol) has on the climate, we need to consider two more things:

- The lifetime of the forcing agent in the atmosphere.
- The radiative efficiency of the forcing agent, i.e. the ability of the forcing agent to cause a change in radiative forcing per unit change in concentration or amount thereof.

Hence the difficulty to compare the impact of different forcing agents without some deeper analysis.

A metric that provides a means to compare the effectiveness of different agents to that of CO₂ is the **Global Warming Potential** (GWP). We can define it as:

$$GWP = \frac{\int_0^{TH} a_x[C_x(t)]dt}{\int_0^{TH} a_r[C_r(t)]dt}$$

where a_x is the radiative efficiency of the forcing agent x , $C_x(t)$ is the time-dependent concentration (decay) of the forcing agent x , a_r is the radiative efficiency of the reference gas (CO₂), $C_r(t)$ is the time-dependent concentration (decay) of the reference gas (CO₂) and TH is the time horizon, or length of time over which we evaluate the impact of the agent relative to the reference. Using CO₂ as the reference gas is simply a convention.

Example: Methane

We assume that we have an initial concentration of methane C_0 that decays exponentially with time (assumption). We say that after time τ the concentration of methane has decreased by a factor of e (63%). We can then calculate the absolute GWP for methane (the numerator in the equation above) as:

$$\text{AGWP}_x = \int_0^{\text{TH}} a_x [C_x(t)] dt = a_x C_{0x} \int_0^{\text{TH}} [e^{-t/\tau}] dt = a_x C_{0x} \tau (1 - e^{-\text{TH}/\tau})$$

We note that CH₄ has a radiative efficiency by **mass** of ~ 72 x that of CO₂ so we write that

$$a_x C_{0x} \approx 72 a_r C_{0r}$$

and CH₄ lifetime is ~ 12 years compared to CO₂ lifetime of ~ 200 years so we can write that the global warming potential for methane is:

$$\text{GWP}_{\text{CH}_4} = 72 \frac{12 [1 - e^{-\text{TH}/12}]}{200 [1 - e^{-\text{TH}/200}]}$$

For reference we can see that in 20 years the GWP for methane is 37x that of CO₂, in 100 years it is 11x and 500 years it is only 4.7x.

This metric will be useful for policy makers to make a first order assessment of the climate impact of different choices (e.g. changes to energy production, agricultural practices, etc.) on different timescales.

7 Climate Timescales and Sensitivity

It is apparent that the climate system has some inertia and that the response to an external forcing is not immediate. In L7 a “thought experiment” is outlined where the Earth is initially assumed to be in equilibrium and then its CO₂ concentration is doubled. In it we see that, as a result of said **thermal inertia**, the Earth requires some time for it to reach a new equilibrium. In the following sections we will try to quantify how long it takes for this new equilibrium to be reached, i.e. what is the **adjustment timescale**.

7.1 The First Law of Thermodynamics Applied to the Climate System

We begin by writing the first law of Thermodynamics in terms of energy:

$$dU = dQ + dW \quad \text{Energy}$$

and we can rewrite this in terms of power ($\frac{d}{dt}$):

$$\frac{dU}{dt} = \frac{dQ}{dt} + \frac{dW}{dt} \quad \text{Power}$$

where U is the internal energy of the system, Q is the heat added to the system and W is the work done **on** the system (contrary to what we have seen in the Thermodynamics course).

We note all the variables are time dependent, so we can write for the control case and the perturbed (2x CO₂) case:

$$\frac{dU_{CTL}(t)}{dt} = \frac{dQ_{CTL}(t)}{dt} + \cancel{\frac{dW_{CTL}(t)}{dt}} \quad \text{and} \quad \frac{dU_P(t)}{dt} = \frac{dQ_P(t)}{dt} + \cancel{\frac{dW_P(t)}{dt}}$$

where the subscript CTL denotes the control case and P denotes the perturbed case. We also note that the work done on/by the Earth is negligible.

Let us also define primed variables to be the differences between the perturbed and control cases, i.e.:

$$U'(t) = U_P(t) - U_{CTL}(t) \quad \text{and} \quad Q'(t) = Q_P(t) - Q_{CTL}(t)$$

it therefore follows from their definitions that

$$\frac{dU'(t)}{dt} = \frac{dQ'(t)}{dt}.$$

We write ΔQ_{ext} is the external (anthropogenic) forcing due to the 2x increase in CO₂ and ΔQ_{int} is the corresponding internal feedback response. It then follows that at some given time t the total heating on the system (Earth) is the sum of both these terms ($\Delta Q_{ext} + \Delta Q_{int}$). We can then write:

$$\frac{dQ'(t)}{dt} = [\Delta Q_{ext}(t) + \Delta Q_{int}(t)] 4\pi (R_E + z_{atm})^2 \quad \because R_E \gg z_{atm}$$

where R_E is the radius of the Earth and z_{atm} is the height of the atmosphere, which is much smaller than R_E and can therefore be neglected.

From equation 3.1 from section 3 we can rewrite $\Delta Q_{int}(t)$ in terms of the feedback parameter:

$$\Delta Q_{int}(t) = \gamma \Delta T_s(t) \implies \Delta T_s(t) = T_{s,P}(t) - T_{s,CTL}(t) = T'_s(t)$$

and combining all of the above we can finally write:

$$\boxed{\frac{dU'(t)}{dt} = (\Delta Q_{ext}(t) + \gamma \Delta T'_s(t)) 4\pi R_E^2}$$

To proceed we will need some assumptions:

- The only feedback process operating is blackbody feedback, i.e. we can rewrite γ as γ_{BB} .
- The biospheric response is manifested in both oceans and land in for the sake of simplicity.
- Temperature change is constant throughout the atmosphere, i.e. assume no lapse rate change (see section 4.2 for details on lapse rate).
- The atmosphere keeps a uniform surface temperature perturbation, i.e. surface temperature perturbation of the land and ocean are the same.

Recalling the definition of heat capacity $U = C\Delta T$, and using sensible values for these we can come up with rough estimates for how much energy the land, ocean and atmosphere can store provided we also have estimates for their mass. Lastly we can also consider the energy stored in the ice sheets, which is in the form of latent heat. Let us write the following:

$$\begin{aligned}\Delta U'_{atm} &= m_A c_A \Delta T'_A(t) \\ \Delta U'_{land} &= m_L c_L \Delta T'_L(t) \\ \Delta U'_{ocean} &= m_{UO} c_O \Delta T'_{UO}(t) + U'_{DO} \\ \Delta U'_{ice} &= l_f m'_{ice}(t)\end{aligned}$$

For detailed computation of the order of magnitude of these terms see L7 slides. From there we will use the fact that heat capacity of the land and atmosphere is about 2 orders of magnitude lower than that of the upper ocean (10^{21} vs 10^{23}) for land/atmosphere vs ocean respectively. We can therefore neglect the former two terms. For the cryosphere, the order of magnitude is significant, its calculation is covered on L7 slides, but we decide to ignore it due to reasons. We are therefore left with only the ocean term so we write:

$$(Q_{ext} + \gamma_{BB} T'_s) 4\pi R_E^2 \approx m_{UO} c_O \frac{dT'_S}{dt} + \frac{dU'_{DO}}{dt}$$

7.2 Adjustment Timescale

We will need to do some further derivation to reach the adjustment timescale. Let's begin by using the boxed equation 7.1 from the previous section and introducing the concept of an 'overturning' circulation strength, or mass exchange rate Ψ (mass/time) between the upper and deep ocean. We can then write:

$$\frac{dU'_{DO}(t)}{dt} = \Psi c_O T'_S(t) - \Psi c_O T'_{DO}(t) = \Psi c_O (T'_S(t) - T'_{DO}(t))$$

where a similar equation can be written for the upper ocean. With this newly defined term and the original equation we can combine them to write:

$$m_{UO} c_O \frac{dT'_S(t)}{dt} = 4\pi R_E^2 (\Delta Q_{ext}(t) - \gamma_{BB} T'_S(t)) - \Psi c_O (T'_S(t) - T'_{DO}(t))$$

and by defining $m_{UO} = A_0 h_0 \rho_0$ where A_0 is the surface area of the ocean, h_0 is the depth of the upper ocean and ρ_0 is the density of the upper ocean, and by dividing by area we can obtain a relation in terms of power per unit area as follows:

$$h_0 \rho_0 c_O \frac{dT'_S(t)}{dt} = \frac{4\pi R_E^2}{A_0} (\Delta Q_{ext}(t) - \gamma_{BB} T'_S(t)) - \frac{\Psi c_O}{A_0} (T'_S(t) - T'_{DO}(t))$$

where the term in **red** is the original forcing and the terms in **blue** are the damping of the change in surface temperature (and therefore upper ocean internal energy) due to the blackbody feedback (first blue term) and the heat exchange between the upper and deep ocean, i.e. oceanic circulation (second blue term).

With this we can define η as the ratio of the damping terms, i.e.:

$$\eta = \frac{\Psi c_O}{4\pi R_E^2 \gamma_{BB}}$$

so it follows that if $\eta \rightarrow 0$ then $\Psi \rightarrow 0$. This is the weak oceanic circulation case.

7.2.1 Weak Oceanic Circulation ($\eta \ll 1$)

In this case we can neglect the second blue term in the above equation and we can write:

$$\frac{dT'_S(t)}{dt} + \frac{4\pi R_E^2 \gamma_{BB}}{m_{UOCO}} T'_S(t) = \frac{4\pi R_E^2 \Delta Q_{ext}(t)}{m_{UOCO}}$$

which is a first order linear differential equation (which we solve by using an integrating factor). We also have some **initial conditions**: $T'_S(0) = 0$ since at $t = 0$ the Earth is in equilibrium and there has not been any change in temperature yet. The solution is then:

$$T'_S(t) = \frac{\Delta Q_{EXT}}{\gamma_{BB}} (1 - e^{t/t_a})$$

where, finally, we encounter the **adjustment timescale** t_a which is defined as:

$$t_a = \frac{m_{UOCO}}{4\pi R_E^2 \gamma_{BB}} \implies T'_S(t_a) = \frac{\Delta Q_{EXT}}{\gamma_{BB}} (1 - e^{-1}) \approx 0.63 \frac{\Delta Q_{EXT}}{\gamma_{BB}} \approx \frac{2\Delta Q_{EXT}}{3\gamma_{BB}}$$

where the last approximation is the one given in the lectures.

It also follows that the equilibrium surface temperature ($T'_{S,eq}$) is reached as time reaches t_{eq}

$$T'_{S,eq} = \frac{\Delta Q_{EXT}}{\gamma_{BB}}$$

which we can see in figure 8.

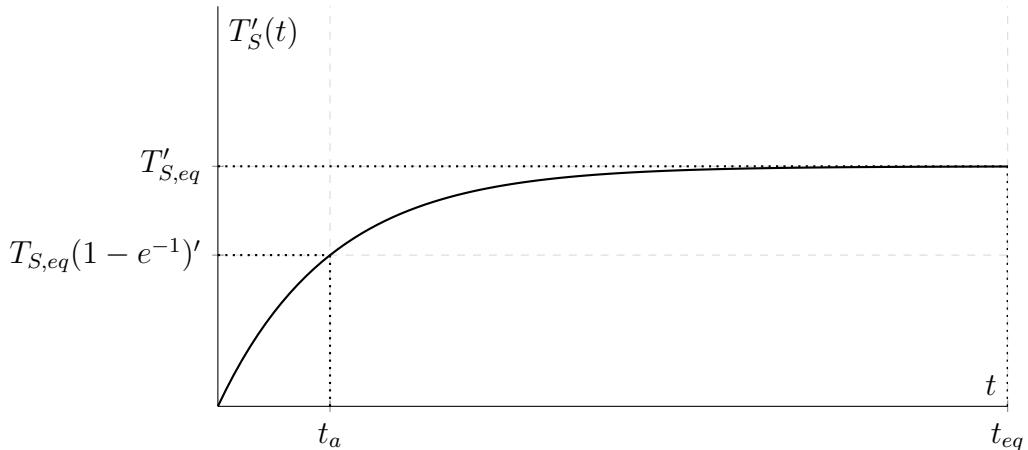


Figure 8: Surface temperature change over time for the weak oceanic circulation

Order-of-Magnitude Estimates:

We have expressions for t_a and $T'_{S,eq}$ as follows:

$$t_a = \frac{m_{UO}c_O}{4\pi R_E^2 \gamma_{BB}} \quad \text{and} \quad T'_{S,eq} = \frac{\Delta Q_{EXT}}{\gamma_{BB}}$$

and we can use these to obtain some order-of-magnitude estimates. We use the following values:

- $m_{UO} = 3.4 \times 10^{19} \text{ kg}$
- $c_O = 4.2 \times 10^3 \text{ J kg}^{-1}\text{K}^{-1}$
- $R_E = 6.4 \times 10^6 \text{ m}$
- $\gamma_{BB} = 4\epsilon'\sigma T_{S,eq}^3 \approx 3.8 \text{ W m}^{-2}\text{K}^{-1}$
- $\Delta Q_{EXT} \approx 4 \text{ W m}^{-2}$

From which we obtain:

$$t_a \approx 2.3 \text{ years} \quad \text{and} \quad T'_{S,eq} \approx 1 \text{ K}$$

7.2.2 Strong Oceanic Circulation ($\eta \gg 1$)

In this case the upper and lower oceanic layers are strongly coupled, so we say that $T'_{DO}(t) \approx T'_S(t)$ and therefore the second blue term in the equation cancels out. Once again, the equation becomes a first order linear DE:

$$\frac{dT'_S(t)}{dt} + \frac{4\pi R_E^2 \gamma_{BB}}{(m_{UO} + m_{DO})c_O} T'_S(t) = \frac{4\pi R_E^2 \Delta Q_{EXT}(t)}{(m_{UO} + m_{DO})c_O}$$

The solution, just as before is exponential and we can write for t_a :

$$t_a = \frac{(m_{UO} + m_{DO})c_O}{4\pi R_E^2 \gamma_{BB}} \approx \frac{m_{DO}c_O}{4\pi R_E^2 \gamma_{BB}} \sim 100 \text{ s years}$$

So we can see that the interaction between the upper and lower oceanic layers has a significant effect on the adjustment timescale, i.e. it is a significant damping factor on the rate of change of $T'_S(t)$. This is shown in figure 9.

7.3 Climate Sensitivity

In the section above (section 7.2) we have assumed a number of things, some of which were not entirely realistic.

One such assumption was the fact that no feedbacks other than blackbody feedback were operating. We therefore used γ_{BB} as the feedback parameter but in fact we know that $\gamma = -\gamma_{BB} + \gamma_{H_2O} + \gamma_{Albedo} + \dots$ so we must rethink our model. We have seen that $t_a \propto \frac{1}{\gamma}$ so it follows that changing γ will change the adjustment time t_a .

We define the **climate sensitivity** as the equilibrium global mean surface temperature change in response to a doubling of CO₂ concentration, i.e. the previously seen $T'_{S,eq}$. To

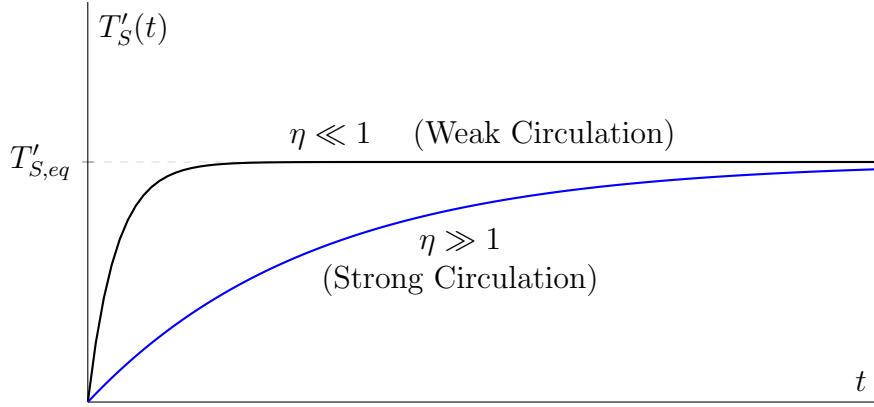


Figure 9: Strong vs weak oceanic circulation surface temperature change over time

improve upon the previous blackbody-only model, we can define the **net feedback parameter** f as follows:

$$f = \frac{\gamma_{H_2O} + \gamma_{Albedo} + \gamma_{Clouds} + \dots}{\gamma_{BB}} = \sum_i \frac{\gamma_i}{\gamma_{BB}}$$

where γ_i is the feedback parameter due to the i -th feedback process we wish to account for. We can now rewrite the climate sensitivity ($T'_{S,eq}$) in terms of f as follows:

$$T'_{S,eq} = \frac{\Delta Q_{EXT}}{\gamma_{BB}} = \boxed{\frac{\Delta Q_{EXT}}{\gamma_{BB}(1-f)}}$$

such that we now only have to consider the net feedback parameter f to gain insight into the **stability** of the climate system.

The following for some values of f follow:

- For negative values of f (i.e. $f < 0$) the initial forcing ΔQ_{EXT} is damped such that the adjustment time reduces and the equilibrium change in surface temperature is reduced. I.e. the equilibrium temperature is still higher than the initial temperature (because the 2x increase in CO₂) is still present, but with a negative f this increase in temperature is less than would be expected from the blackbody-only model.
- For $f = 0$ we essentially have the blackbody-only model.
- For positive values of f (i.e. $f > 0$) the initial forcing ΔQ_{EXT} is amplified such that the adjustment time increases and the equilibrium change in surface temperature is increased.

Finally, we notice that a value of $f = 1$ would mean that the climate system is unable to reach an equilibrium, i.e. the planet is unable to ever radiate enough energy to space to counteract the effect of the original forcing and all the operating feedbacks. This is called a **runaway greenhouse effect**.

8 Observing Our Climate

As we have seen in previous sections, signals of climate change are quite small in section 7 we see that a 2x in CO₂ concentration “only” leads to a few degrees temperature difference. It is of very high importance therefore to be able to measure these changes accurately and precisely (see glossary for definitions) as to be able to detect anthropogenic signals and distinguish them from **natural variability**.

8.1 Examples of Potential Issues: In-Situ and Ground Based Measurements

8.1.1 The Urban Heat Island Effect

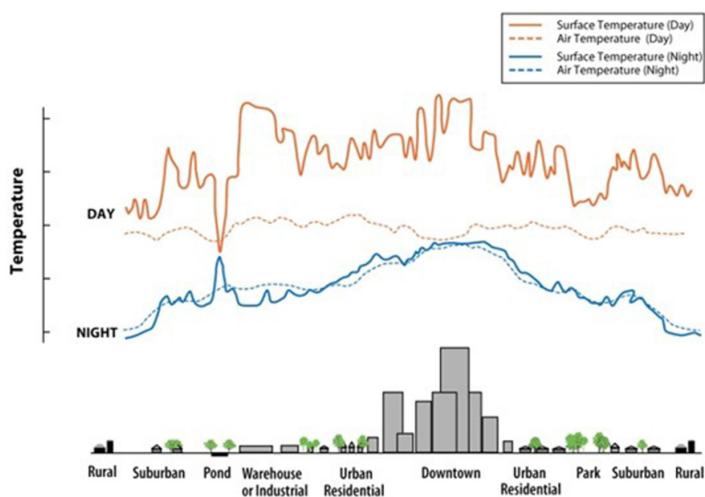


Figure 10: Urban Heat Island Effect.

Figure 10 shows a schematic of the **urban heat island effect**. This is a phenomenon where urban areas are warmer than their rural surroundings. This is especially true at night. Some reasons for this are:

- **Albedo:** Urban areas have a lower albedo than rural areas. This means that they absorb more SW radiation and therefore heat up more. This is obvious when we consider that some key urban materials like asphalt have an albedo much lower of that of vegetation.
- **Thermal Inertia:** Urban areas have a lower thermal inertia than rural areas. This means that they heat up more quickly during the day and cool down more quickly during the night. The fact that they heat up more during the day means that they release more heat during the night, further Increasing temperatures at when the sun does not shine.
- **Anthropogenic Heat:** Urban areas have a higher concentration of anthropogenic heat sources (cars, air conditioning, heating, etc.) than rural areas.
- **Evapotranspiration:** Urban areas have a lower evapotranspiration rate than rural areas. This means that less energy is used to evaporate water and more energy is used to heat the air. This is the case due to the lack of vegetation and bodies of water (ponds, lakes, etc.) in urban areas.

We can define the **Urban Heat Island Index** (somehow abbreviated UHI) as the maximum temperature difference between the urban and rural areas as a function of time:

$$\boxed{\text{UHI}(t) = \max(T_{\text{urban}}(t) - T_{\text{rural}}(t))}$$

which as previously mentioned is usually larger at night. In fact during the day it is only usually the air that is closest to the ground that is different. We can also consider the following **surface energy balance**:

$$(R_S^\downarrow - R_S^\uparrow) + (R_L^\downarrow - R_L^\uparrow) = H + E + Q$$

where the R_S terms represent the net downward SW radiation at the surface, the R_L terms represent the net downward LW radiation at the surface, H is the sensible heat flux¹, E is the latent heat flux and Q is the ground heat transfer. It therefore follows that each of right hand side terms are affected due to urban characteristics.

It is therefore obvious that having measurement stations in/near urban areas will lead to a bias in the measurements. This is a problem because most of the **long term climate records** are based on measurements from urban areas. Without corrections, warming trends would be amplified due to urbanisation.

8.1.2 Changes in Sensor Design/Instrument Characteristics

A clear example of how changes in how measurements are taken can lead to biases in the data is the change in the design of **sea surface temperature** (SST) measurements. There are a number of ways in which SST measurements can be taken including floating buoys, bucket measurements, hull contact sensors and ship engine intake. It is clear, especially for the latter two measurement methods, that the readings will have a bias. Such biased measurement techniques are being phased out as we move towards more accurate and precise methods so once again it is crucial to make corrections when comparing new and old data. See L9 for lower-level details.

Another example of this is the change in the design of global **radiosonde** network. Radiosondes are instruments that are attached to weather balloons and measure atmospheric parameters (temperature and humidity) as they ascend through the atmosphere. The first measurements date back to the early 1900s so it is clear that the design of the instruments has changed significantly since then. Furthermore, different regions of the world have used different types of sensors so once more it is crucial to make corrections when comparing new and old data and when comparing data from different regions when making a globally homogeneous record.

Examples of what can cause discrepancies across different sensors include:

- **Time constant:** The time it takes for the sensor to reach a new temperature after a change in temperature. This arises due to the sensor's Thermal Inertia.
- **Material composition:** Different materials have different Emissivity and Absorptivity characteristics, meaning that they will absorb and emit different amounts of radiation.
- **Sensor location:** The length of the string attaching the sensor to the balloon can affect the readings. This is because the balloons tend to have warmer temperatures than the environment around them.

¹Sensible heat flux means that is heat that can be “sensed”, or measured, usually with a thermometer. Latent heat flux in contrast cannot be measured with a thermometer.

8.2 Surface Temperature Trends and the Impact of Satellite Observations

Despite the issues mentioned above, after applying the necessary corrections, we can still see a warming trend of about 1K over the last 100 years and this warming is more accentuated over land than over the oceans as per section 7.

Satellite observations have become a standard practice since the late 1970s for gathering data on the state of the atmosphere and surface, including temperature and humidity. These observations offer significant advancements over traditional surface and in-situ measurements, evident in their higher spatial resolution and the availability of a more comprehensive and global sea-surface temperature (SST) record starting from the late 1970s.

8.3 Space-Based Observations

In the previous section we outline some of the advantages of space-based observations when compared to “legacy” means of climate observation. However, space-based observations come with challenges and limitations of their own.

There are two main types of space-based sensors:

- **Passive Sensors:** These sensors measure the LW radiation emitted by the Earth and backscattered light from the Sun. These are the most common type of sensors and have been used since the beginning of the space-based observation era to measure surface and atmospheric temperature.

These are low-power devices so they are able to scan large portions of the Earth.

- **Active Sensors:** These newer sensors emit their own radiation and measure the amount of radiation that is reflected back to them (LIDAR, RADAR). These kind of sensors are particularly good at measuring clouds and aerosols and their vertical structure.

These are high-power devices so they are only able to scan small portions of the Earth at once.

The climate variable that a sensor is measuring will be dependent on the wavelength examined by it. Always remember that the only kind of interaction between the Earth and a satellite will be through **photon exchange**.

Up to now we have mostly considered **irradiance** I_λ : the amount of energy per unit area per unit wavelength emitted or received from all directions such that at the TOA we have:

$$\text{OLR} = \int I_\lambda d\lambda$$

but most satellite sensors measure **radiance** L_λ : the amount of energy per unit area per unit wavelength per unit solid angle received from a particular direction. Newer satellites are able to measure radiance in multiple directions. Furthermore, it follows that if we are interested in these instruments giving us insight into the temperature of the surface or the atmosphere that the light captured by these must be in the LW part of the spectrum.

Aside on Radiance vs Irradiance:

Let's consider an observational instrument in space aimed at Earth, designed to measure Earth's reflected and emitted energy.

Irradiance:

Suppose the instrument is equipped with a sensor that measures the total power incident on it per unit area from Earth, integrating over the entire field of view. This sensor does not differentiate between the various regions on Earth's surface it's viewing - it's simply recording the total power from all these regions per unit area. The measured quantity is the irradiance at the sensor's location, arising from the totality of the Earth scene within its field of view.

Radiance:

Now, consider a different instrument on the same platform, but this one is a hyperspectral imaging device - a bit like a camera sensor that can capture many different wavelengths and has many pixels. This instrument is designed to measure the power received in each of its pixels per unit area per unit solid angle. Each pixel corresponds to a specific area on Earth's surface, and the instrument measures the power coming from that specific area and direction. For each pixel, it quantifies the power received in each of a multitude of narrow spectral bands, thereby providing a spectrum for each pixel.

The quantity it measures is radiance. It is a directional quantity, associated with a specific location (the corresponding surface area) and a specific direction (the direction from that area to the instrument).

For a given location on Earth's surface, the radiance varies with the direction. For example, if the location is a forest, the radiance in the direction of the sun may be quite different from the radiance in a direction away from the sun, due to the scattering properties of the trees.

Therefore, while irradiance gives a measure of the total power per unit area received from all directions, radiance provides a detailed picture of the power per unit area received from each specific direction. This allows for a more nuanced understanding of how light interacts with Earth's surface and atmosphere.

8.4 Obtaining Temperature Information from Space and Schwarzschild's Equation

To obtain temperature information from space we need to consider the following:

- The interaction between the Earth's emitted LW radiation and the atmosphere.
- How to describe these interactions mathematically.
- Translate this to temperature measurements from space.

8.4.1 Interaction Between LW Radiation and the Atmosphere

There are three main ways in which LW radiation interacts with the atmosphere: **absorption**, **emission** and **scattering**. We can neglect scattering in the LW in times of clear-sky condi-

tions. This is not the case if there are clouds, but in such a case we cannot measure the surface temperature anyway.

We can begin by considering absorption by a layer of atmosphere dz . As per the Beer-Lambert Law, we have the following:

$$dL_{\lambda\text{abs}} = -L_{\lambda}\rho_a\kappa_{\lambda}^a dz$$

where $dL_{\lambda\text{abs}}$ is the change in radiance due to absorption, L_{λ} is the incident radiance at the base, ρ_a is the density of the absorbing material and κ_{λ}^a is the mass absorption coefficient.

Secondly, we consider the emission by a similar layer of atmosphere dz . As per Kirchoff's Law (LTE) we have the following:

$$dL_{\lambda\text{em}} = B_{\lambda}(T_z)\rho_a\kappa_{\lambda}^a dz$$

where $dL_{\lambda\text{em}}$ is the radience emitted by the layer, $B_{\lambda}(T_z)$ is the **Plank Function** at the temperature of the layer T_z and $\rho_a\kappa_{\lambda}^a$ is the Emissivity (equal to the Absorptivity).

We can combine these two equations to obtain the following:

$$dL_{\lambda} = -[L_{\lambda} + B_{\lambda}(T_z)]\rho_a\kappa_{\lambda}^a dz$$

and bring in the concept of optical depth τ_{λ} do do the following derivation:

$$\begin{aligned} \tau_{\lambda} &= \int \rho_a\kappa_{\lambda}^a dz \implies d\tau_{\lambda} = \rho_a\kappa_{\lambda}^a dz \\ dL_{\lambda} &= -[L_{\lambda} - B_{\lambda}(T_z)]\rho_a\kappa_{\lambda}^a dz \\ \frac{dL_{\lambda}}{d\tau_{\lambda}} + L_{\lambda} &= B_{\lambda}(T_z) \end{aligned}$$

which is a first order ODE that we can solve for L_{λ} using an integrating factor and the following boundary conditions (see L10 for more details):

$$L_{\lambda}(z_{\text{space}}) = L_{\lambda}(0)e^{-\tau_{\lambda}} + \int_0^{\tau_{\lambda}} B_{\lambda}(T_z)e^{-(\tau_{\lambda}-\tau'_{\lambda})} d\tau'_{\lambda}$$

This is the **Schwarzschild Equation**. The term in orange is the radiance seen by the satellite, the term in blue is the radiance emitted by the surface that reaches the satellite despite the absorption by the atmosphere and the term in red is the radiance emitted by the atmosphere that reaches the satellite.

We can rewrite the above equation in terms of **Transmissivity** where we use the following notation to write transmissivity of the atmosphere from 0 to z_{space} : $t_{\lambda}(0, z_{\text{space}}) = e^{-\tau_{\lambda}}$:

$$t_{\lambda}(z', z_{\text{space}}) = e^{-(\tau_{\lambda}-\tau'_{\lambda})} \implies \frac{dt_{\lambda}(z', z_{\text{space}})}{d\tau'_{\lambda}} = e^{-(\tau_{\lambda}-\tau'_{\lambda})}$$

and we obtain the Schwarzschild Equation in terms of transmissivity:

$$L_{\lambda}(z_{\text{space}}) = L_{\lambda}(0)t_{\lambda}(0, z_{\text{space}}) + \int_{t_{\lambda}(0, z_{\text{space}})}^1 B_{\lambda}(T_z) dt'_{\lambda}(z', z_{\text{space}})$$

Remember that we make the assumption that **the atmosphere is in LTE** and that there is **no scattering**.

8.4.2 Weighting Functions

When we attempt to retrieve temperature information from space, we will not use the Schwarzschild Equation as written above, but rather we transform it into the following in terms of altitude z :

$$L_\lambda(z_{\text{space}}) = L_\lambda(0)t_\lambda(0, z_{\text{space}}) + \int_{z=0}^{z_{\text{space}}} B_\lambda(T_z) \frac{dt'_\lambda(z', z_{\text{space}})}{dz} dz$$

where the term in red is the **weighting function** $K_\lambda(z)$. Note that the weighting function could be defined in terms of any other variable so long as it is monotonic² with height.

The concept of a weighting function is a useful one because it allows us to understand where from the atmosphere a particular wavelength is coming from as measured by a satellite.

It follows that by scanning a number of different wavelengths we are able to obtain a vertical profile of the atmosphere for the particular feature we are interested in³ e.g. temperature, humidity, etc. Finally then, the higher the spectral resolution (i.e. the narrower the range of wavelengths the satellite measures) the more vertical information the satellite can obtain.

8.5 Issues With Satellite Based Records

As with any other kind of measurement, satellite based measurements are not without their own issues. Some of these include:

- **Accuracy of Calibration:** Instruments must be calibrated against a known reference or standard prior to launch to space as to convert a measurement to the required quantity. The **transfer of calibration** from ground to space is often unstable.
- **Length of Missions and Consistency of Instrumentation:** A typical mission will last for about 3-5 years. This means that instruments are replaced somewhat regularly, and although they are replaced with newer and potentially slightly different equipment as to exploit new technologies, which can lead to inconsistencies in the data.
- **Orbit Decay:** The orbit of a satellite will decay over time due to atmospheric drag. This means that the satellite's altitude changes leading to changes in the measured signal.
- **Gaps in Data:** If launches are delayed, or if the satellite in orbit malfunctions and there are no others that can take over, then there will be gaps in the data.
- **Inversions:** Inversions are the process of transforming satellite measurements into the required quantity. Like we have seen above, these inversions are “models” and as such have their limitations and assumptions. It therefore follows that different models will yield different insight with the same data. (There is some extra detail in the last few slides of L10, mainly examples).

²Monotonic means that one variable is either always increasing or always decreasing with respect to the other variable. It also ensures that there is a one-to-one correspondence between variables. E.g. pressure and height.

³This requires a constant gas density, so usually the CO₂ band is used

9 Modelling Our Climate

Throughout the previous sections, we have discussed a number of climate models with varying levels of complexity. They are all based on the fundamental principle of energy conservation and equilibrium, i.e. energy input = energy output for a planet in equilibrium, known as energy balance models (EBM). In this section we will have a look at some of these models and explore different ways in which we can apply this principle to make more nuanced models.

9.1 Energy Balance Models

9.1.1 The Zero and One-Dimensional Energy Balance Models

The simplest energy balance model is the zero-dimensional energy balance model. We have covered this extensively in sections 1 and 2. It is the model of photon energy exchange where the Earth gets all its energy from the Sun and radiates it back into space (through albedo, blackbody radiation etc.).

Within this model we explored the **zero-layer** version, i.e. the no-atmosphere model where we only account for the energy coming from the sun, albedo and blackbody radiation from the Earth's surface. This was the energy balance equation:

$$(1 - \alpha)\text{TSI} = \sigma T_s^4$$

Also in the aforementioned lectures is the **one-layer** version, i.e. the atmosphere is treated as a single layer. In such a model we also account for the emissivity of the atmosphere and its absorptivity (which we stated are equal in value for a system in LTE). This was the energy balance equation:

$$(1 - \alpha)\text{TSI} = (1 - \epsilon_a)\sigma T_s^4 + \epsilon_a\sigma T_a^4$$

and it also lead to the introduction of the greenhouse effect. For more details see section 2.

We can now extend this model to make it what is known as the **1-D EBM**. In this model we introduce latitudinal dependence. This is an obvious first choice for independent variable since it is apparent that the energy balance is going to be different in the poles than in the equator.⁴. For now we will only consider the mean annual values for the energy balance, see figure 11.

As we see in figure 11, the energy balance is different in the poles than it is in the equator (when averaged over the year). The relevance of this is that this will set the main circulation patterns of energy (both atmospheric and oceanic).

Let's express this model mathematically. We begin by writing the energy balance equation:

$$\frac{(1 - \alpha)}{4}\text{TSI} = \epsilon'\sigma T_s^4$$

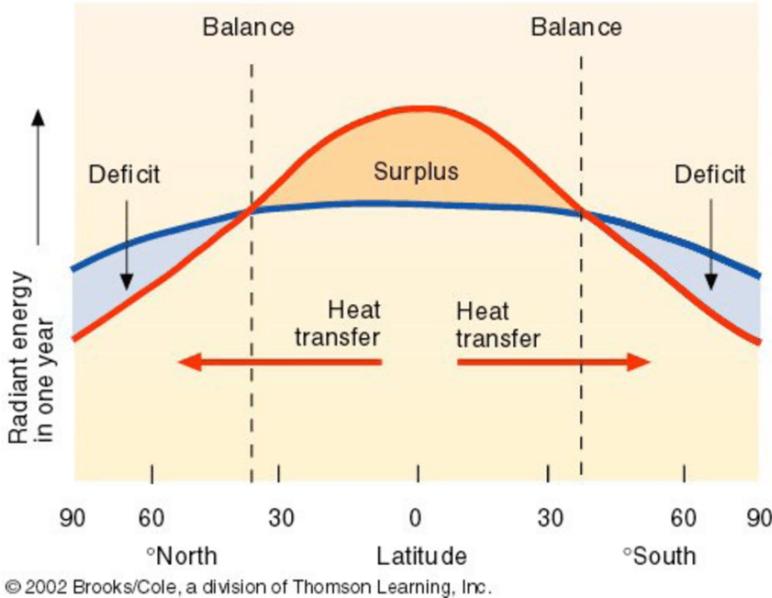
where ϵ' is the effective emissivity of the Earth (see 3.2). It follows that for a perturbation from equilibrium

$$C \frac{dT_s}{dt} = \frac{(1 - \alpha)}{4}\text{TSI} - \epsilon'\sigma T_s^4$$

where C is the heat capacity per square meter. For a short range of values for T_s we can linearise as follows and substitute into the equation:

$$\epsilon'\sigma T_s^4 \approx A + BT_s \quad \Rightarrow \quad C \frac{dT_s}{dt} \approx \frac{(1 - \alpha)}{4}\text{TSI} - (A + BT_s)$$

⁴We could also add time-dependency, since it is apparent that this also has a significant effect



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Figure 11: The 1-D Energy Balance Model. The energy balance is different in the poles than in the equator.

where A and B are constants. We can now “slice” the Earth into small latitudinal bands and consider the energy balance in each band i :

$$C_i \frac{dT_{s,i}}{dt} \approx \frac{(1 - \alpha_i)}{4} \text{TSI}_i - A - BT_{s,i}$$

and finally we can include a term for the lateral heat transport (F) to complete the model:

$$C_i \frac{dT_{s,i}}{dt} + F(T_{s,i} - T_s) \approx \frac{(1 - \alpha_i)}{4} \text{TSI}_i - A - BT_{s,i}$$

where F is in units of $\text{Wm}^{-2}\text{K}^{-1}$. The one-dimensional energy balance model **must be solved numerically**.

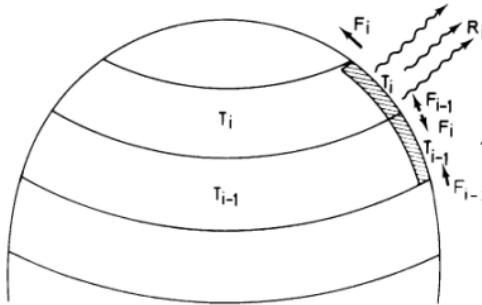


Figure 12: Diagram of the 1D EMB.

9.1.2 Coupled General Circulation Models (CGCMs)

These are the most complex and complete models available, so called “state-of-the-art”. These are the kind of models used by the IPCC to make predictions about the future. In the course we do not go too much into detail about these models, precisely due to their complexity.

In these models, both the atmosphere and the ocean are divided into a grid of boxes. The atmosphere is further divided into layers of altitude. These systems are modelled as a set of coupled non-linear ordinary differential equations that govern the evolution of the systems from a fluid dynamics perspective (in the atmosphere and ocean) while always ensuring conservation of momentum, mass and energy.

Models have evolved significantly over time, with more and more layers of complexity being added as computing power available increases and new inter-dependencies within the climate system are identified (i.e. the science evolves). Figure 13 shows the evolution of CGCMs over time with more components being added as time goes on.

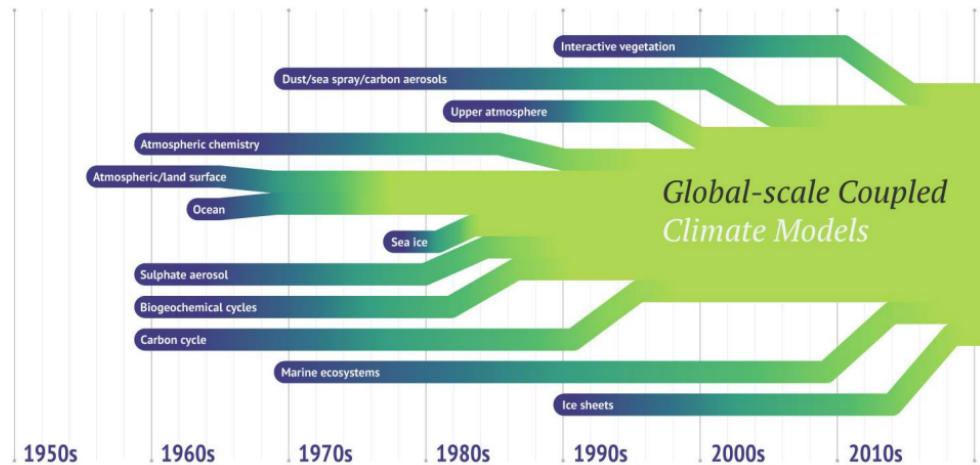


Figure 13: Evolution of CGCMs over time.

9.1.3 Earth System Models

These are the absolute most advanced models available. They are essentially CGCMs but not only simulate physical processes (oceans, land, atmosphere) but also biogeochemical processes (most notably the carbon cycle) and the way they interact with each other. This allows to introduce some additional feedbacks into the model that are not present in CGCMs such as how warming may affect the ability of forests to absorb CO₂, or how changes in ocean circulation may alter the ocean's CO₂ uptake.

9.2 Model Parametrisation

In climate modelling, certain physical process cannot be modelled in a per-cell basis due to their scale being too small, e.g. convection, cloud formation and evolution and others. To model these, we use **parametrisation**.

In parametrisation, we find simplified representations or approximations of sub-grid phenomena, expressed in terms of the larger-scale variables that the model does directly resolve. They are often based on empirical evidence from observations or more detailed process-based studies, or on theoretical understanding from physical laws. Within parametrisation techniques we find two main types:

Diagnostic Scheme: A diagnostic scheme determines the outcome (e.g., the presence and properties of clouds) directly from the current state of the large-scale variables, with no memory of past states. In other words, given the same current state, a diagnostic scheme will always

produce the same outcome.

Prognostic Scheme: A prognostic scheme is more advanced, it includes some form of temporal evolution or memory, meaning that the outcome can depend on past states as well as the current state. This allows the model to represent processes that have some inherent time-dependence or persistence, such as the life cycle of clouds or the accumulation of snow. Given the same current state, a prognostic scheme could produce different outcomes depending on the prior sequence of states.

The main idea to remember between the two is that prognostic schemes have memory (time dependence) while diagnostic schemes do not.

Also worth remembering is that clouds are the main source of uncertainty in climate models by far, analytical solutions become impossible so with every approximation comes error (error in frequency of occurrence, error in amount and properties and error in location and timing).

9.3 Costs of Climate Modelling

Climate models are very expensive to run, both in terms of computing power and time.

- Atmosphere is the most expensive component of the model to run and is typically the bottle neck of climate models (radiation and chemistry schemes are very computationally intensive)
- Timestep in Earth System Model worlds is typically 20 minutes for dynamics and 1h for radiation
- 1 week of very high performance computer time yields about 40 years of simulated climate. When we consider that we want to simulate 1000 or 2000 years of climate per model, and include many of them in an ensemble for an IPCC report we can see how this becomes a very expensive and time consuming process.
- There is about 1 million ‘basic’ variables only to model oceans and atmosphere. This is very memory consuming

10 Climate Policy and Implications

It is clear that there is an anthropogenic impact on the climate as per all of the information presented in the previous sections. The question now is what can be done about it such that humanity can continue to enjoy prosperity.

Precisely because anthropogenic climate change is believed to be a threat to said prosperity, there has been international recognition of the need to implement policies to mitigate the effects and the scale of dangerous climate change. The most clear example of this is the 2015 Paris Agreement, which was signed by 200 countries and sets out the following goals:

- To limit the increase in global surface temperature by <2K above pre-industrial revolution levels by the end of 2100.
- To achieve “net-zero” emissions by the second half of the 21st century, allowing for carbon capture technology to be used to offset any remaining emissions.
- To set up a “global stocktake” to assess whether the goals are being met by individual countries. This is also called the **intended nationally determined contributions** (INDCs). The first such stocktake will be in 2023.

The Paris Agreement is complementary to the UN’s 17 Sustainable Development Goals (SDGs), which are a set of goals to be achieved by 2030. See Figure 14.



Figure 14: The 17 Sustainable Development Goals

10.1 How to Achieve the Climate Goals

The above goals are ambitious, so if we wish to achieve them we have to analyse what the existing patterns of emissions are and how they need to be changed to achieve the goals.

Figure 15 shows us the breakdown of global CO₂ emissions by sector. We clearly see that if we wish to limit the anthropogenic impact of climate change, we need to focus on the energy sector.

Global greenhouse gas emissions by sector

This is shown for the year 2016 – global greenhouse gas emissions were 49.4 billion tonnes CO₂eq.

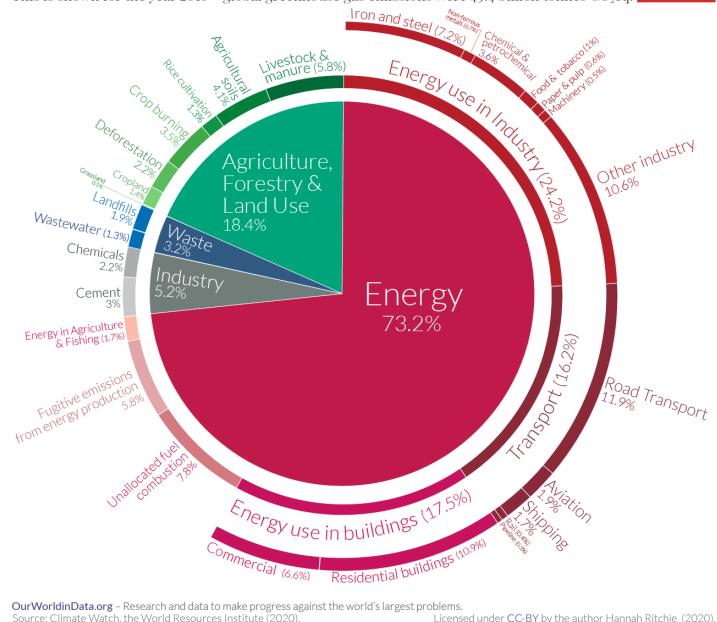
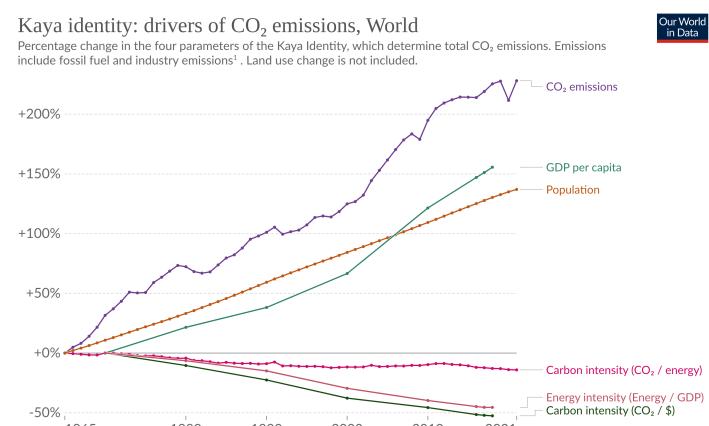


Figure 15: Global CO₂ emissions by sector

The **Kaya Identity** allows us to relate the total emissions to four key factors: population, GDP per capita, energy intensity and carbon intensity:

Total Emissions F (kg)	=	Population P	×	GDP per capita GDPP (\$ pp)	×	Energy Intensity EI (kWh/\$)	×	Carbon Intensity CI (kg/kWh)
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We see that in the past decades the increase in emissions (F) is largely due to the population increase as well as the increase in GDP per capita. By contrast, the energy intensity and carbon intensity have decreased, though to a smaller extent. See Figure 16. We also note that these individual factors change from country to country (for more details see L12 slides 9 onwards).



¹ Fossil emissions: Fossil emissions measure the quantity of carbon dioxide (CO₂) emitted from the burning of fossil fuels, and directly from industrial processes such as cement and steel production. Fossil CO₂ includes emissions from coal, oil, gas, flaring, cement, steel, and other industrial processes. Fossil emissions do not include land use change, deforestation, soils, or vegetation.

Figure 16: The Components of the Kaya Identity

10.2 SSPs and RCPs

Representative Concentration Pathways (RCPs), are scenarios of future concentrations of greenhouse gases and other forcings. RCPs were used in the (IPCC's) Fifth Assessment Report.

The RCPs are named after their radiative forcing target level for the year 2100, expressed in (W/m²). There are four RCPs:

- **RCP 2.6:** This is a "peak and decline" scenario, where greenhouse gas emissions peak between 2010 and 2020, with substantial declines thereafter. By 2100, the additional radiative forcing is limited to 2.6 W/m².
- **RCP 4.5 and RCP 6:** These are stabilization scenarios where total radiative forcing is stabilized before 2100 by applying a range of technologies and strategies for reducing greenhouse gas emissions. The additional radiative forcing in 2100 is 4.5 W/m² and 6 W/m², respectively.
- **RCP 8.5:** This is a rising radiative forcing pathway leading to 8.5 W/m² by 2100. This scenario assumes that emissions continue to rise throughout the 21st century.

These pathways provide inputs for climate model simulations to project future climate change. It's important to note that they're not predictions or forecasts, but are plausible scenarios based on specific assumptions about future socio-economic development, technological changes, and policy interventions.

The RPC approach has been extensively criticised by placing too much emphasis on the most pessimistic of outcomes. As a result thereof, the IPCC has introduced a new set of scenarios for the Sixth Assessment Report, called the **Shared Socioeconomic Pathways** (SSPs).

The SSPs link population growth and movement, technological development, energy production and use, land use and cost to future greenhouse gas and aerosol emissions into 5 different "choices" or scenarios that are as follows (directly quoted):

- **SSP1: Taking the Green Road.** World shifts towards a more sustainable path, common solutions to health and educational issues, economic growth places emphasis on human well-being. Greater equality results.
- **SSP2: Middle of the Road.** Social, economic and technological trends follow historical patterns leading to uneven growth. There is slow progress towards sustainable development goals.
- **SSP3: Regional Rivalry - A Rocky Road.** Nations put their own interests first at the expense of wider development and international action on sustainable development goals (e.g. health, education, climate).
- **SSP4: Inequality - A Road Divided.** Increasing inequality both across and within nations. Effectively the rich get richer and the poor get poorer. Conflict and unrest become increasingly common.
- **SSP5: Fossil-fuelled development - Taking the Highway.** Growth in innovation and competitive (but integrated) markets allows rapid technological progress. Strong investment in health and education. Development is fuelled by fossil fuel exploitation. Geo-engineering is a viable option (see section 13)

The different SSPs lead to different emission levels (with SSP1 being the lowest and SSP5 being the highest). Aerosol emissions are reduced in all SSPs. It is important to note that none of these scenarios are able to match the Paris Agreement goals without additional implementation of mitigation strategies (carbon capture, geoengineering etc.).

For some lower-level details on the SSPs (GDP, GDP per capita, total energy demand, impact on the global mean surface temperature) see L12 slides 18 onwards or Riahi *et al.* 2017.

11 Harnessing Wind Energy

As has become apparent in section 10, there is a significant need to reduce the carbon emissions from the energy sector. One of the ways to achieve this is to substitute fossil fuels as an energy source for renewable sources of energy. While sun and wind power are thought to be the ones most likely to grow the fastest in the next decades, this section will focus on wind power and how it can be harnessed.

11.1 Extracting Power from the Wind

The main idea about wind power is that, by using a device (wind turbine) we are able to extract energy from the wind (kinetic energy) and transform it into some other form of energy that is more useful to us (electrical power).

Let's consider a model in which air is an incompressible fluid (poor assumption), with density ρ the unperturbed wind is initially travelling with speed V_1 and swept cross-sectional area S_1 and after passing through the wind turbine it is travelling with speed V_2 and swept cross-sectional area S_2 . See Figure 17.

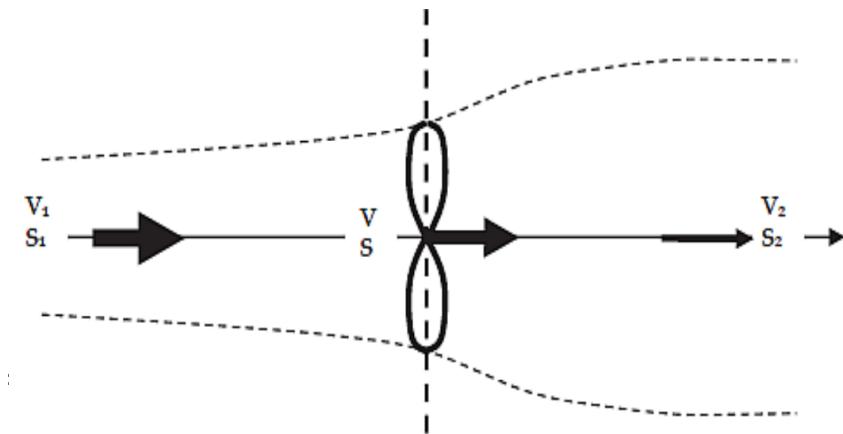


Figure 17: Wind Turbine Model

With this in mind we can note a few things. Firstly, the fact that the turbine extracts kinetic energy suggests that $V_2 < V_1$. Secondly, as per conservation of mass flow, we can write that

$$\frac{dm}{dt} = \rho V_1 S_1 = \rho V_2 S_2 = \rho SV = \text{constant}$$

and from it we can derive the power we are able to extract from the wind:

$$\begin{aligned} \text{Force applied by wind } F &= m \frac{dV}{dt} = \frac{dm}{dt} \Delta V = \rho SV(V_1 - V_2) \\ \text{Power from wind } P &= F \frac{dx}{dt} = \rho SV^2(V_1 - V_2) \end{aligned}$$

Similarly, we can write the power in terms of change of kinetic energy:

$$P = \frac{1}{2} \frac{dm}{dt} (V_1^2 - V_2^2) = \frac{1}{2} \rho SV (V_1^2 - V_2^2)$$

and equating the two expressions for power we get:

$$\begin{aligned} \frac{1}{2} \rho SV (V_1^2 - V_2^2) &= \rho SV^2 (V_1 - V_2) \\ \frac{1}{2} (V_1 + V_2) (\cancel{V_1 - V_2}) &= V (\cancel{V_1 - V_2}) \end{aligned}$$

$$V = \frac{1}{2}(V_1 + V_2)$$

$$P = \frac{1}{4}\rho S(V_1^2 - V_2^2)(V_1 + V_2)$$

so we have obtained a formula for the total available power to be extracted from the wind.

It is evident however, that due to limitations similar to those we encountered in Carnot cycles in thermodynamics we will not be able to harness 100% of all available power. It therefore follows that we will be interested in how efficient our wind turbine is, and therefore can define the **performance coefficient** C_p . To do that we firstly introduce the concept of the kinetic power content, W i.e. the power contained in a unit mass of air moving with initial speed V_1 :

$$W = \frac{1}{2} \frac{dm}{dt} V_1^2 = \frac{1}{2} \rho S V_1^3$$

and so we define the performance coefficient as the power extracted from the wind turbine divided by the kinetic power the wind had in the first place:

$$C_p = \frac{P}{W} = \frac{\frac{1}{4}\rho S(V_1^2 - V_2^2)(V_1 + V_2)}{\frac{1}{2}\rho S V_1^3} = \frac{(V_1^2 - V_2^2)(V_1 + V_2)}{2V_1^3}$$

to which we can introduce the interference factor $b = V_2/V_1$ and make some substitutions:

$$P = \frac{1}{4}\rho S V_1^3 (1 - b^2)(1 + b)$$

$$C_p = \frac{1}{2}(1 - b^2)(1 + b)$$

and finally we can plot the performance coefficient as a function of the interference factor (see figure 18).

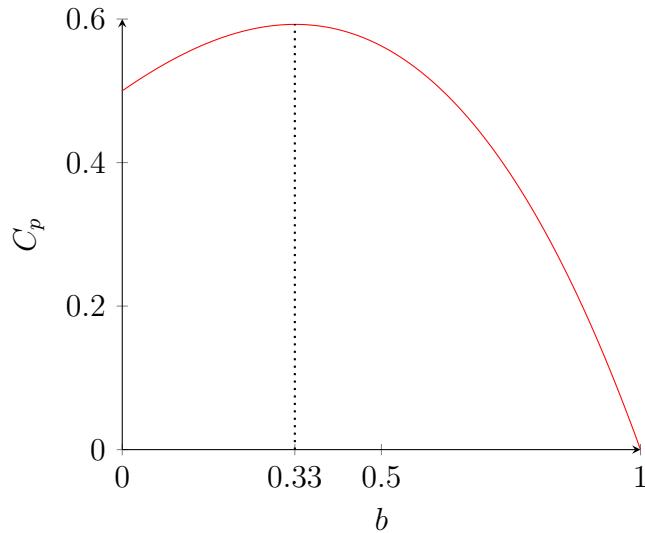


Figure 18: Performance coefficient as a function of interference factor

From the plot we can see that the maximum performance coefficient is $C_p = 0.593$ and it occurs at $b = 1/3$. This is the **Betz Limit** and it is the maximum theoretical power fraction that can be extracted from a wind stream. Analytically we can write:

$$\frac{dC_p}{db} = \frac{1}{2}(1 - 3b)(1 + b) \implies b = \frac{1}{3} \quad \therefore V_2 = \frac{1}{3}V_1$$

which using the above definition for P we get:

$$P = \frac{16}{27} \frac{\rho}{2} V_1^3 S$$

and assuming a circular swept area of diameter D i.e. $S = \pi D^2/4$ we get:

$$P = \frac{16}{27} \frac{\rho}{2} V_1^3 \frac{\pi D^2}{4}$$

In reality, just like with the Carnot cycle, it is not realistic to obtain this maximum efficiency due to frictional losses in the rotor, drag due to blade surface roughness and other mechanical imperfections. In practice, the maximum efficiency is around 40-45%.

11.2 Wind-Speed Distributions

In order to be able to estimate how much power will be able to be extracted from the wind it is important to know the wind speed distribution as to place wind turbines in the most optimal of places.

We typically approximate the wind-speed distribution $f(V)$ as a Rayleigh distribution:

$$f(V)_{\text{Rayleigh}} = \frac{V}{c^2} e^{-(V/c)^2} \quad 0 \leq V < \infty$$

and in some situations, the Weibull distribution might be more appropriate:

$$f(V)_{\text{Weibull}} = \frac{k}{c} \left(\frac{V}{c} \right)^{k-1} e^{-(V/c)^k} \quad 0 \leq V < \infty$$

where c and k are distribution parameters. The Weibull goes to a Rayleigh when $k = 2$.

We can therefore use the Rayleigh distribution to estimate the average extractable power. For a turbine operating in a flow of instantaneous speed V the extractable power P is:

$$P = C_p \frac{V^2}{2} \frac{dm}{dt} = \frac{1}{2} C_p \rho S V^3$$

and as per what we said before, we can find a mean for the power by using the mean of the distribution of wind speeds:

$$\begin{aligned} \overline{P(V)} &= C_p \int_{V_{\min}}^{V_{\max}} P(V) f(V)_{\text{Rayleigh}} dV \\ &= C_p \int_0^\infty \frac{\rho S V^3}{2} f(V)_{\text{Rayleigh}} dV \\ &= \frac{C_p \rho S}{2} \int_0^\infty V^3 f(V)_{\text{Rayleigh}} dV \\ \boxed{\overline{P(V)} = \frac{C_p \rho S}{2} \overline{V^3}} \end{aligned}$$

Note that $\overline{V^3}$ is not the same as \overline{V}^3 . In this case, we use $\overline{V^3} = c^3 3\sqrt{\pi}/4$ for the equations above.

It therefore follows that at the Beltz Limit:

$$\overline{P(V)}_{\max} = \frac{16}{27} \frac{\rho}{2} \frac{\pi D^2}{4} \frac{6}{\pi} \overline{V}^3 = \rho \left(\frac{2}{3} D \right)^2 \overline{V}^3$$

Another factor to bear in mind is the fact that the wind speed is not constant on the vertical axis: it increases with height. This is in fact the reason why wind turbines have been getting taller and larger in diameter over the years.

A good model for the vertical distribution of speeds is a logarithmic profile as follows:

$$V(z) = k \ln \left(\frac{z}{z_0} \right)$$

where k is a constant and z_0 is the **roughness length** - a measure of the height above the surface at which the windspeed theoretically becomes zero. Roughness length is indeed determined by the roughness of the surface, i.e. how much of an impediment does it pose to the wind travelling above it.

Surface	Roughness Length (m)
Sand	0.0001-0.001
Open Sea	0.005
Grass	0.01-0.1
Suburban	0.2-0.4
Urban	0.35-0.8
Forest	0.9-1

Table 1: Roughness Lengths

11.3 Issues with Wind Power

There are a number of issues with wind power that have to be considered. Here is a non-exhaustive list:

- Wind supply is not constant, meaning that there will be times when generation is needed but is not able to be provided, and times of excess generation. This has to be addressed by having a backup power source, or by having a way to store the excess energy.
- They are noisy and can be an eyesore.
- Power transmission/distribution is an issue. Wind farms are typically located in remote areas, meaning that the power has to be transmitted over long distances, leading to potential losses.
- There is a non-zero carbon impact of building and replacement of wind turbines. An average lifespan for a wind turbine is 10-20 years.
- Large scale deployment of wind turbines create wake effects, i.e. the performance of a turbine is affected when it is behind another one. Turbine separation must be 5-10x rotor diameter.

12 Solar Energy

In previous sections we have seen the importance of an increase in renewable energy generation in order to reach climate goals. In the previous section we concentrated on wind power, in this one we will look at solar power.

Near the beginning of the course we noted that TSI at TOA is $\sim 1360 \text{ W/m}^2$, which if averaged over the whole Earth is $\sim 240 \text{ W/m}^2$ when accounting for albedo, and **in total 120 PW** (global electricity consumption is $\sim 3 \text{ TW}$, i.e. much smaller). For these figures, we assumed that there is no atmospheric absorption or scattering in SW. Remember of course that this is not quite the case - SW is mostly unaffected by the atmosphere but not fully. See fig 2

Let's examine this assumption a bit more closely. As per section 2.2, we had a definition of optical depth, in that case just for an absorbing atmosphere.

It is apparent as per fig 2 that the atmosphere is not just absorbing but also scattering, especially in the SW. As a result we will modify the definition for optical depth to replace the mass absorption coefficient for the **mass extinction coefficient** $\kappa_{e\lambda}$:

$$\tau_\lambda = \int_0^z \kappa_{e\lambda}(z) \rho_a(z) dz \quad \text{where} \quad \kappa_{e\lambda} = \kappa_{a\lambda} + \kappa_{s\lambda}$$

so the solar radiation reaching the surface becomes:

$$I_{\text{surface}} = I_{\text{TOA}} e^{-\tau_\lambda}$$

so it becomes apparent that **path length** and therefore how much atmosphere is interacting with the sunlight before reaching the surface is important in determining how much of it will be available at the surface.

Before moving on, we must first define some concepts that will allow us to progress.

12.1 Some Definitions

To account for path length, for solar energy applications we will often encounter the concept of **air mass** (AM). The concept of air mass is a way to quantify the amount of atmospheric matter (such as air molecules, aerosols, etc.) that sunlight must traverse before it reaches the Earth's surface. This path length of sunlight through the atmosphere is directly related to the amount of scattering and absorption the sunlight undergoes, which impacts its intensity by the time it reaches the surface.

The AM is defined as the ratio of the actual path length that sunlight travels through the Earth's atmosphere relative to the path length it would travel if the sun were directly overhead, i.e. at zenith. At zenith, sunlight travels the shortest possible path through the atmosphere, and so this path length serves as our baseline (AM=1) for defining air mass. We define AM mathematically as:

$$\text{AM} = \sec(\theta_z) \implies I_{\text{surface}} = I_{\text{TOA}} e^{-\tau_\lambda \sec(\theta_z)}$$

note that an air mass of 0 implies extra-terrestrial, TOA solar irradiance at the surface, i.e. no atmosphere.

A useful number for air mass is 1.5, which is the average air mass at a latitude in a "temperate climate" $\sim 42^\circ$ latitude (e.g. Barcelona). This is useful since population tends to be closer to latitudes like these rather than the equator or poles.

In the above equation we used θ_z to denote the **solar zenith angle**, this is the angle between the direction of the sun and the vertical (zenith direction). When the sun is directly overhead, $\theta_z = 0$ and the air mass is 1. As the sun moves away from zenith, θ_z increases, and the sunlight has to traverse a longer path through the atmosphere, thus increasing the air mass.

Total solar radiation at the surface can be broken down into two components: **direct** and **diffuse** radiation.

Direct Radiation: This is the portion of solar radiation that comes straight from the direction of the Sun without undergoing any kind of scattering or reflection. It follows a direct path from the Sun to the Earth's surface. This is of special interest to technologies like concentrating photovoltaics (CPV), which work best with direct sunlight (they use concentrating optics so they track the sun).

Diffuse Radiation: This is the portion of solar radiation that has been scattered in various directions by molecules, aerosols, and clouds in the Earth's atmosphere. Instead of coming from a single direction, like the Sun, diffuse radiation comes from multiple directions. This scattered light reaches the Earth's surface from different angles and is important for standard photovoltaic (PV) panels which can absorb sunlight from various directions (these tend to be fixed so don't track the sun).

The following two terms are not crucial to know, but can be useful for a better understanding of the topic and its literature.

The term **global radiation** can sometimes cause confusion, as it might suggest something related to the entire globe. However, "global" radiation simply refers to the total amount of solar radiation reaching a specific location on Earth's surface, i.e., it's the sum of direct and diffuse radiation at said point.

Another term that might appear is **Direct Normal Irradiance** (DNI). This refers to the amount of solar radiation received per unit area by a surface that is always held perpendicular (or normal) to the rays that come in a straight line from the direction of the sun. This is particularly relevant for solar concentrator systems, which aim to maximize their exposure to direct radiation.

12.2 Photovoltaic Fundamentals

A number of further concepts are introduced in Lecture 14b, most of which are not crucial to know for the exam. Here follows a brief summary of the most important ones.

Solar Materials: Different materials have different uses when it comes to harnessing solar energy. For solar **thermal** applications, metals or black solids are often used because they are excellent at absorbing sunlight and converting it into heat. For solar chemical applications, plants are used in processes like photosynthesis, where sunlight is used to convert carbon dioxide and water into glucose and oxygen. Sometimes other chemicals are used for solar chemical processes as well, such as in the production of hydrogen from water using photoelectrochemical (PEC) cells.

Photovoltaics: Photovoltaic (PV) materials are used in solar panels to convert sunlight directly into electricity. The key material in PV cells is a semiconductor, often silicon, which is 'doped' with impurities to create a so-called p-n junction with properties favorable for solar energy conversion. The band-gap of this semiconductor material is crucial - it must be suitable to absorb the energy from incoming solar photons, which can then excite electrons across the band-gap, creating an electric charge.

The band gap is the energy difference between the valence band (where electrons are normally located) and the conduction band (a higher energy level where electrons can move freely). When a photon with energy equal to or greater than the band gap hits the PV cell, it can excite an electron from the valence band to the conduction band, creating an electron-hole pair. The ‘size’ of the band gap greatly affects the performance of the PV cell. A large band gap will be excited by high energy photons and will produce a high voltage, as they leave behind a hole in the valence band and there is a large difference between the two (potential difference). However, a large band gap will not be excited by as many electrons (the lower energy ones, higher wavelength) so the current of such cell will be lower.

Photogenerated Charges: Once the solar photons have been absorbed and the electrons are excited across the band-gap, these photogenerated charges (electron-hole pairs) need to be separated to prevent them from simply recombining. The p-n junction and an associated electric field help in this charge separation. Asymmetric contacts made to an external circuit then allow these charges to flow as an electric current.

Power and Efficiency: The power produced by a PV cell is the product of the photovoltage (the voltage generated by the photovoltaic effect) and the photocurrent (the current generated by the separated charges). The maximum power point of a cell is where the product of the photovoltage and photocurrent is greatest. The power conversion efficiency of a PV cell, meanwhile, is the ratio of the power (or power per unit area) that is converted into useful work (electricity), divided by the incident power (or irradiance) from the sun. Consistency in units is crucial when calculating efficiency.

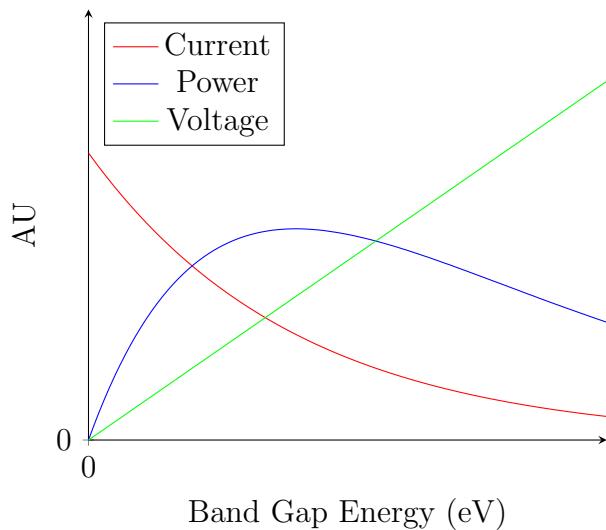


Figure 19: Current, voltage and power curves for different band gaps. Note that these are estimates and not to scale.

12.3 PV Technology

In the lecture a number of PV technologies, each with their advantages/drawbacks were introduced. Here follows a summary of the most important ones.

Single-Junction Silicon Cells: These are the most commonly used type of PV cell and are typically composed of monocrystalline or polycrystalline silicon. They rely on a single

semiconductor material (silicon) with a specific band gap. This restricts the range of light wavelengths that can be absorbed and turned into electrical energy, hence limiting the maximum theoretical efficiency to about 33.7% under standard test conditions⁵ (known as the Shockley-Queisser limit). In practice, the efficiencies achieved are usually around 15-20%, with the highest efficiencies nearing 26%.

Multi-Junction Cells: These PV cells are composed of multiple layers, each made from different semiconductor materials with different band gaps. This allows the cell to absorb a broader range of light wavelengths, enhancing its efficiency. Each layer in the multi-junction cell is designed to absorb a specific range of the solar spectrum, allowing it to capture more of the incoming solar energy than a single-junction cell could. As a result, multi-junction cells can achieve efficiencies approaching 50%. However, they are also more complex and expensive to manufacture.

Alternative Material-Based Cells: These are a diverse group of emerging PV technologies that use materials other than traditional semiconductors. This includes organic (plastic) solar cells, dye-sensitized solar cells, and perovskite solar cells. These technologies offer the potential for lower costs and more flexible applications, but they currently face challenges related to efficiency, stability, and scaling.

⁵Said standard test conditions try to replicate what the cell would experience under 1.5 SM, i.e. 1000 W/m² and a particular frequency spectrum. This spectrum can be found online but all that is worth knowing is that it tries to replicate real-world conditions and that it is normalised for 1000 W/m².

13 Geoengineering Our Climate

For this course we will take the following definition of geoengineering:

“The deliberate, large-scale intervention in the Earth’s natural systems to counteract climate change.”

Two main types of geoengineering are considered:

- **Carbon Dioxide Removal (CDR):** The removal of CO₂ from the atmosphere in an attempt to minimise the atmosphere’s overall greenhouse effect and hence reduce global mean surface temperature. This increases the OLR side of the energy-balance equation. Examples include afforestation, ocean fertilisation and direct air capture.
- **Solar Radiation Management (SRM):** The reflection of incoming solar radiation away from the Earth in an attempt to reduce the amount of energy absorbed by the Earth’s surface. This reduces the absorbed solar radiation side of the energy-balance equation. Examples include stratospheric aerosol injection, low cloud albedo enhancement and space-based reflectors.

Before we discuss the various methods of geoengineering, we must introduce a concept that we will have to bear in mind throughout the rest of this section - **EATS**. This stands for the following:

- **Effectiveness:** Is the proposed method going to have a significant, measurable impact on the climate system?
- **Affordability:** Is the proposed method likely to be affordable?
- **Timeliness:** Is the proposed method likely to be available in time to prevent dangerous climate change? How mature is this technology?
- **Safety:** Is the proposed method likely to have any negative side-effects?

And lastly (not in this mnemonic) we should also consider legality, and feasibility, international cooperation, public perception etc.

13.1 Carbon Dioxide Removal

In this section we will discuss various examples of CDR.

Bio-Energy with carbon capture and storage (BECCS): This is a combination of two technologies - bio-energy and CCS. The idea is to grow biomass (trees), burn it to generate energy and then capture the CO₂ emitted during combustion through CCS techniques. The captured CO₂ is then transported and stored underground or under the sea. This process therefore acts as a net **carbon sink** - it removes CO₂ from the atmosphere and stores it underground. This is sometimes referred to as “negative emissions”.

However, it is worth noting that implementing this is not without its challenges. Mainly, this requires a vast amount of land to grow the biomass on, which negatively impacts food security and biodiversity. Furthermore, the time-scale is not particularly favourable and there are significant monetary costs involved.

Enhanced Rock Weathering: This is a climate change mitigation strategy that accelerates the natural process of weathering to remove carbon dioxide from the atmosphere. It involves pulverizing rocks rich in silicate minerals, like basalt or olivine, and spreading them over large areas. The increased surface area of the powdered rock allows for faster chemical reactions with CO₂, locking it away in stable forms such as bicarbonate ions. This not only reduces atmospheric CO₂ levels but also replenishes soil nutrients for crops and promotes plant growth, which further absorbs CO₂ (feedback). Furthermore, when eventually these are washed away by rivers into the ocean these are the “right kind” of carbonates that help reduce ocean acidification. Despite its potential, the method poses challenges, including the energy-intensive process of mining and grinding rocks and the potential reduction of albedo due to the dark rocks.

13.2 Solar Radiation Management

Marine Cloud Brightening: This is a proposed form of SRM that leverages the Twomey effect (see Section 6.2), i.e. the indirect effect. The idea is to spray aerosols into the atmosphere which, in areas of marine stratocumulus (this occurs especially to the west of continents) will increase the number of cloud droplets and hence increase the cloud albedo significantly. This is suggested to be done by having a fleet of ships that pump sea salt into the atmosphere. This will increase the amount of solar radiation reflected back into space, reduce the amount of solar radiation absorbed by the Earth’s surface and therefore reduce global mean surface temperature.

However, this method is not without its challenges. It is not clear whether this system is really scalable. Secondly, there might be hidden carbon costs (having a fleet of ships pumping sea salt into the atmosphere is not exactly carbon neutral).

Glossary

Absorptivity The ratio of the energy absorbed by a surface to the energy incident on it. It is a dimensionless quantity. It goes from 0 to 1. 4, 26, 29, 46

Accuracy The degree to which the result of a measurement, calculation, or specification conforms to the correct value or a standard. 46

Adjustment Timescale The time it takes for the climate system to reach a new equilibrium temperature after a change in forcing. See section 7.2 for more details. 46

Aerosol Small liquid drops or particulates suspended in the atmosphere. They can be natural or anthropogenic in origin and can be made up of many different constituents. They can be emitted directly into the atmosphere (primary) or formed in the atmosphere as a result of chemical processes (secondary). See section 6.2 for more details. 46

Band Gap The energy difference between the valence band (where electrons are normally located) and the conduction band (a higher energy level where electrons can move freely). See section 12.2 for more details.. 46

Beer-Lambert Law The law that describes the attenuation of light through a medium. See section 8.4.1 for more details. 29, 46

Betz Limit The maximum possible efficiency of a wind turbine, ~59%. 46

Climate Feedback Parameter A measure of how much the climate system responds to a change in some surface temperature change. A measure of how much energy is trapped or lost as a result of this change in surface temperature. See 3.1. 46

Climate Sensitivity The equilibrium global mean surface temperature change in response to a doubling of CO₂ concentration. See section 7.3 for more details. 46

Dissolved Inorganic Carbon (DIC) The sum of dissolved carbon dioxide, carbonic acid, bicarbonate and carbonate. 46

EBM Energy Balance Model. A model that describes the temperature of a planet as a function of the energy it receives from the Sun and the energy it radiates back into space. See section 9.1 for more details. 31, 46

Emissivity The ratio of the energy radiated by a surface to the energy radiated by a black body at the same temperature. It is a dimensionless quantity. It goes from 0 to 1. 4–6, 26, 29, 46

Global Warming Potential A metric that provides a means to compare the effectiveness of different agents to that of CO₂. See section 6.3 for more details. 46

Gross Primary Productivity (GPP) The total amount of carbon fixed by photosynthesis in an ecosystem per unit time. 46

Intended Nationally Determined Contributions The climate actions that countries intend to take under the Paris Agreement. They are not legally binding but it is a legal obligation for the countries to submit an NDC every 5 years. 46

IPCC Intergovernmental Panel on Climate Change. A scientific and intergovernmental body under the auspices of the United Nations, set up at the request of member governments, dedicated to the task of providing the world with an objective, scientific view of climate change and its political and economic impacts. It does not produce its own research, rather it gathers insight from the scientific community and summarises it in its Assessment Reports. 36, 46

Irradiance The amount of power per unit area incident on a surface. It is measured in Wm^{-2} and for this course it will always be power from EM radiation. 46

Kaya Identity An equation that expresses the total emissions of CO₂ as the product of four factors: population, GDP per capita, energy intensity and carbon intensity. See section 10.1 for more details. 46

Kirchoff's Law The law that states that the emissivity of a body is equal to its absorptivity at a given wavelength and temperature so long as it is in local thermodynamic equilibrium. 29, 46

Lapse Rate The negative rate of change of temperature with height. It is measured in Km^{-1} . See section 4.2 for more details. 46

LTE Local Thermodynamic Equilibrium - the state of a system in which all parts of the system are at the same temperature and there are no net flows of energy between different parts of the system. 29, 31, 46

LW Long wave. We will use this to refer to long wave EM radiation. 4–6, 8, 11, 12, 26–28, 46

Net Feedback Parameter The sum of all feedback parameters due to all feedback processes operating in the climate system. See section 7.3 for more details. 46

Net Primary Productivity (NPP) The GPP minus the amount of carbon released by respiration i.e. $NPP = GPP - R$. 46

Ocean Biological Carbon Pump The process by which carbon is transported from the surface ocean to the deep ocean. See section 5.4.2. 46

OLR Outgoing Long Wave Radiation. The amount of LW radiation emitted by the Earth. It is measured in Wm^{-2} . 5–7, 11, 27, 46

Optical Depth The measure of how transparent a medium is to electromagnetic radiation. It is a dimensionless quantity. See section 2.2 for more details. 46

Performance Coefficient The ratio of the power extracted by the turbine to the power available in the wind. See section 11.1 for details. 46

Photosynthetically Active Radiation (PAR) The spectral range of solar radiation from 400 to 700 nm that photosynthetic organisms are able to use in the process of photosynthesis. See figure 6. 46

Planetary Albedo The fraction of solar radiation that is reflected back into space. It is a dimensionless quantity. It goes from 0 to 1. See section 1.3 for more details. 12, 46

Precision The degree to which repeated measurements under unchanged conditions show the same results. 46

PV Photovoltaic: Conversion of light into electricity using solar cells.. 43, 46

Radiative Efficiency The ability of a forcing agent to cause a change in radiative forcing per unit change in concentration or amount thereof. 46

Radiative Forcing The change in net irradiance at the TOA due to a change in some external factor. It is measured in Wm^{-2} . 46

Radiosonde An instrument that is attached to weather balloons and measure atmospheric parameters (temperature and humidity) as they ascend through the atmosphere. 46

Rayleigh Distribution A distribution that is often used to model wind speed. See section 11.2 for details. 46

Representative Concentration Pathways A set of four greenhouse gas concentration trajectories adopted by the IPCC for its fifth Assessment Report. They are used as input for climate models to project future climate change and its impacts. See section 10.2 for more details. 46

Roughness Length A measure of the height above the surface at which the windspeed theoretically becomes zero. See section 11.2 for details and table 1 for some example values. 46

Runaway Greenhouse Effect A theoretical scenario in which the climate system is unable to reach an equilibrium as a response to a forcing, i.e. the planet is unable to ever radiate enough energy to space to counteract the effect of the original forcing and all the operating feedbacks. See the end of section 7.3 for more details. 46

Schwarzschild Equation The equation that describes the radiance seen by a satellite as a function of the radiance emitted by the surface and the radiance emitted by the atmosphere. See section 8.4.1 for more details. 46

Shared Socioeconomic Pathways A set of five narratives that describe alternative future developments in the global society, demographics, economy, energy and land use systems. They are used as input for climate models to project future climate change and its impacts. See section 10.2 for more details. 46

SM Solar Mass: The mass of the sun, used as a unit of mass in astronomy.. 46

SST Sea Surface Temperature. 26, 27, 46

Sustainable Development Goals A list of 17 goals set by the United Nations in 2015 to be achieved by 2030. They are a call for action by all countries to promote prosperity while protecting the planet. 46

SW Short wave. We will use this to refer to short wave EM radiation. 4–6, 12, 25, 26, 42, 46

Thermal Inertia The property of a material to resist changes in temperature. The higher the thermal inertia, the longer it takes for the material to reach a new equilibrium temperature. 26, 46

TOA Top of Atmosphere. This is the top of the atmosphere, where the atmosphere ends and space begins. 3–7, 11, 27, 42, 46

Transmissivity The ratio of the energy transmitted by a surface to the energy incident on it. It is a dimensionless quantity. It goes from 0 to 1. 4, 6, 29, 46

TSI (α_p) Total Solar Irradiance. It is the amount of solar irradiance emitted by the Sun. The sum total of photons emitted by it, converted to Wm^{-2} . 3, 5–7, 42, 46

UHI Urban Heat Island Effect. See section 8.1.1 for details. 46

Weighting Function The function that describes the contribution of a particular layer of the atmosphere to the radiance seen by a satellite. See section 8.4.2 for more details. 46

Well-Mixed Greenhouse Gas A gas that is well-mixed in the atmosphere (troposphere), i.e. its concentration does not vary significantly with geographical location or height. See section 6.1 for more details. 46

Zenith Angle The angle between the sun and the zenith (the point directly overhead). See section 12.1 for more details. 46

References

- [1] Brian J Soden and Isaac M Held. An assessment of climate feedbacks in coupled ocean–atmosphere models. *Journal of Climate*, 19(14):3354–3360, 2006.