Lecture 20. Overview of Climate system

1. Weather/Climate

Weather is the meteorological state of the atmosphere that we experience instantaneously at our particular location.

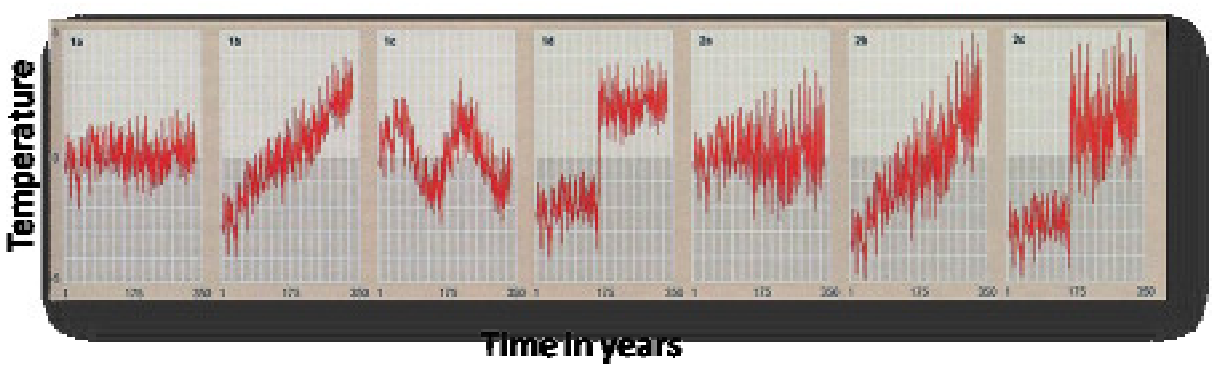
Climate is the mean behavior of the weather over some appropriate averaging time in a specific location.

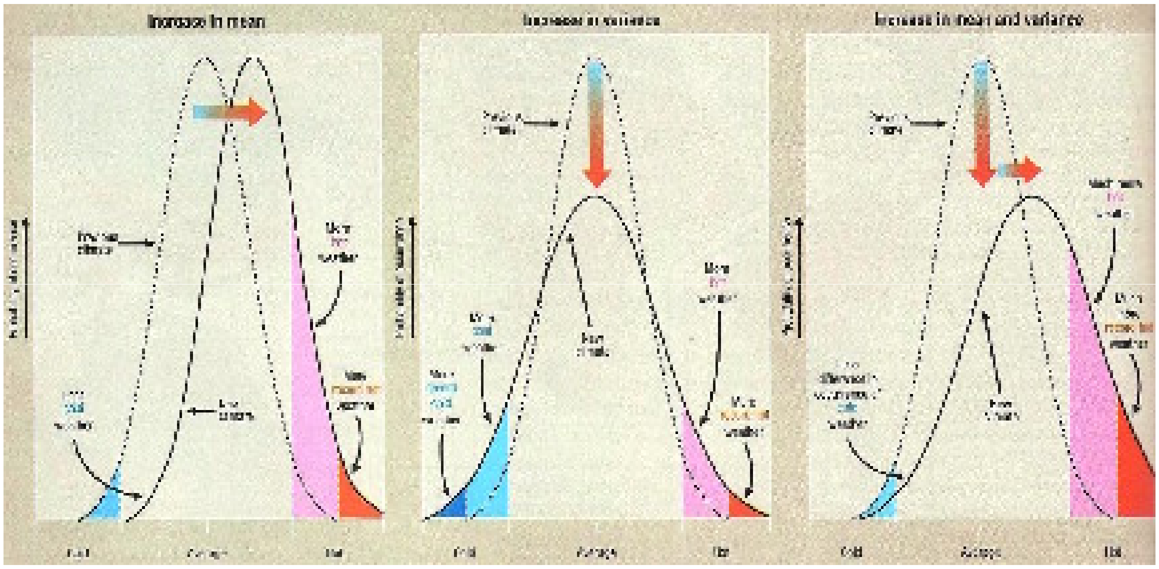
- Note that the definition of climate is flexible for the period of averaging. A traditional averaging time for defining climate is 30 years.
- The most fundamental climate parameter is the planetary (global) average temperature.

NOTE: Global mean surface temperature is the most stable climate parameter, which varies less than a few tenths of a degree over centuries. The atmosphere, land and ocean have an enormous capacity to store heat. Therefore they maintain the stable average temperature despite fluctuations in the global heat balance.

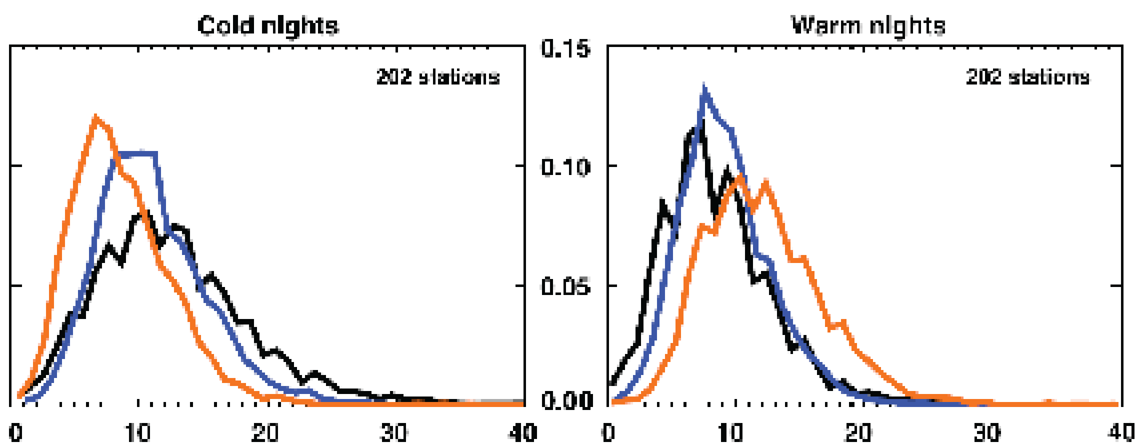
Climate variability involves fluctuation about the mean

Climate change involves change of the mean climate variables. These changes are generally related to overall warming or cooling of the globe.





Example:
Annual probability distribution functions for temperature indices for 202 global stations with at least 80% complete data between 1901 and 2003 for three time periods: 1901 to 1950 (black), 1951 to 1978 (blue) and 1979 to 2003 (red). The x-axis represents the percentage of time during the year when the indicators were below the 10th percentile for cold nights (left) or above the 90th percentile for warm nights (right). IPCC WG1 CH3*.*



Climate stability is the long-term reliability of the climate.
Radiative Equibrium: Application of first law of thermodynamics for planet with net solar radiation Q_{TOA} and outgoing longwave radiation I_{TOA} at the top of atmosphere (TOA) yields

$$\frac{\partial E}{\partial t} = c_p \frac{\partial T_E}{\partial t} = Q_{TOA} + I_{TOA}$$

In a stable climate, $c_p \frac{\partial T_E}{\partial t} = 0$, so $Q_{TOA} = I_{TOA}$ on long timescales. This condition is called Planetary Radiative Equilibrium.
With a planetary albedo Rs, the net shortwave radiation is

$$Q_{TOA} = \pi R_E^2 (1 - R_s) F_0,$$

with $R_E = 6371 \times 10^3$ [km] the Earth's radius, and the net longwave is

$$I_{TOA} = 4\pi\sigma T_E^4$$

with $\sigma = 5.67 \times 10^{-8}$ W.m^{-2}.K^{-4} is the Stefan-Boltzmann constant.

Planetary Energy Balance

Energy emitted by Earth = Energy absorbed by Earth

$$\sigma T_E^4 (4\pi R_E^2) = F_0 \pi R_E^2 (1 - R)$$

$$\begin{aligned}
\sigma T_E^4 &= \frac{F_0}{4}(1 - R\ ) \\
&= \frac{1365\,[W.m^{-2}]}{4} \times (1 - R\ ) \\
&= 341[W.m^{-2}] \times (1 - R\ ) \\
&= 239[W.m^{-2}]
\end{aligned}$$

with $R \approx 0.3$ is the global mean planetary albedo, $R_E = 6371 \times 10^3$ km is the Earth's radius, and $\sigma = 5.67 \times 10^{-8}$ W.m^{-2}.K^{-4} is the Stefan-Boltzmann constant.

Earth's blackbody temperature: $T_E = \left(\frac{239[W.m^{-2}]}{5.67 \times 10^{-8}\,[W.m^{-2}.K^{-4}]} \right)^{\frac{1}{4}} = 255K = -18^0 C$

However the global mean Earth's surface temperature T_s=288 K=15^{o}C. The difference is due to greenhouse effect.

For Earth's surface:

$$\sigma T_s^4 (4\pi R_E^2) = F_0 \pi R_E^2 (1 - R) + \sigma T_E^4 (4\pi R_E^2)$$

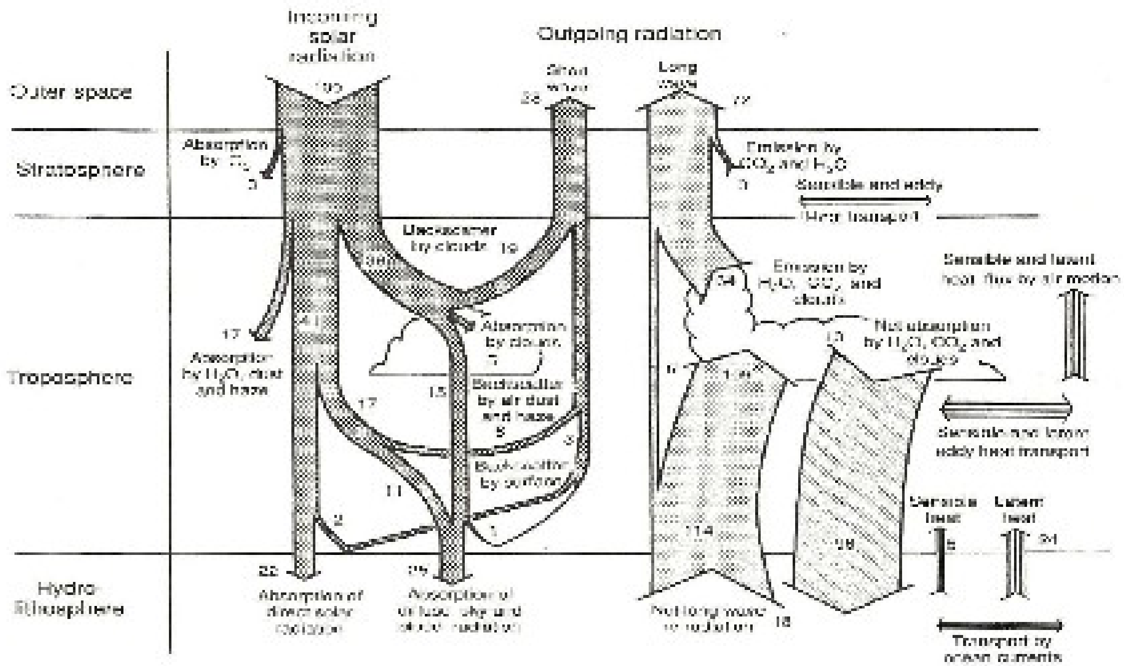
$$\begin{aligned}
T_s &= \left(4 \times 10^9 + T_E^4\right)^{\frac{1}{4}} \\
&= 301K = 28^0 C
\end{aligned}$$

Now, the estimated surface temperature is more than 10^{o}C higher than observed. This is due in part by aerosol scattering.
By assuming a purely scattering aerosol layer of optical thickness τ =0.16, we got

$$\sigma T_s^4 (4\pi R_E^2) = F_0 \pi R_E^2 (1 - R) \exp(-\tau) + \sigma T_E^4 (4\pi R_E^2)$$

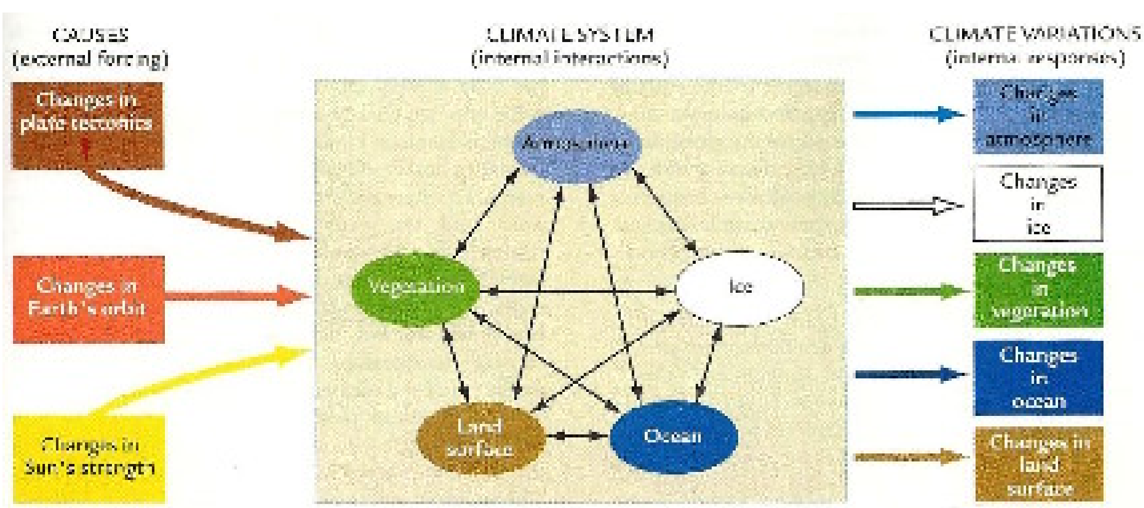
$$\begin{aligned}
T_s &= \left(3.4 \times 10^9 + T_E^4\right)^{\frac{1}{4}} \\
&= 296K = 22^0 C
\end{aligned}$$

which is a closer to observation. To estimate more precisely T_s, we have to consider the climate is a complex system of energy reservoirs joined together (coupled) by energy transfer processes.



<u>Major energy reservoirs of the climate system:</u>

i. **Atmosphere (gases, aerosols, and clouds);**
ii. **Ocean;**
iii. **Land (including biosphere)**
iv. **Ice sheet and sea-ice**

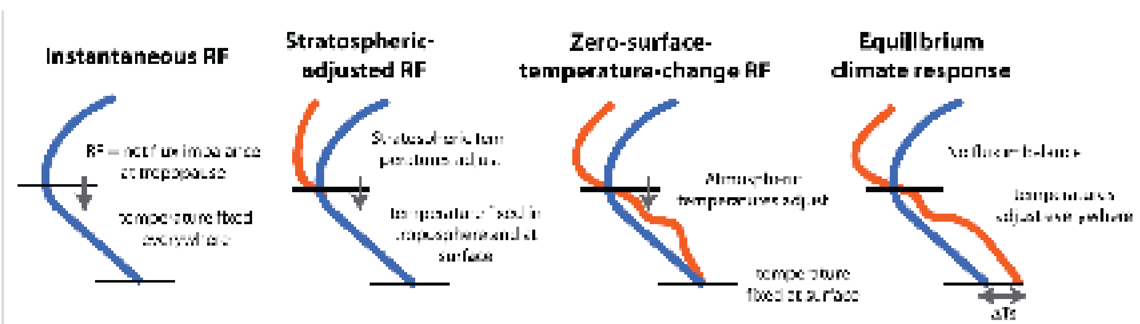


Climate forcing is a change imposed on the climate system that has the potential to alter global temperature.

Examples: a change in solar radiation incident on the Earth is a natural climate forcing; change in atmospheric CO2 abundance due to fossil fuel burning is an anthropogenic forcing.

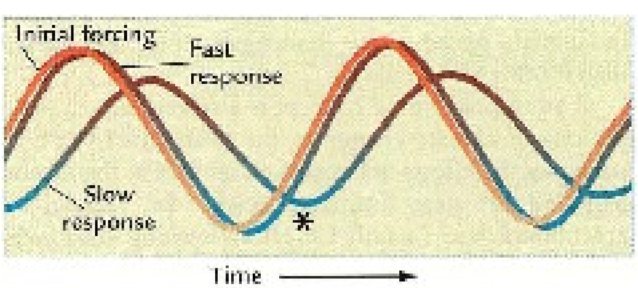
IPCC defines radiative forcing as *the change in net (down minus up) irradiance (solar plus longwave; in W m–2) at the tropopause after allowing for stratospheric temperatures to readjust to radiative equilibrium, but with surface and tropospheric temperatures and state held fixed at the unperturbed values*. Radiative forcing is used to assess and compare the anthropogenic and natural drivers of climate change.

Schematic comparing RF calculation methodologies. Radiative forcing, defined as the net flux imbalance at the tropopause, is shown by an arrow. The horizontal lines represent the surface (lower line) and tropopause (upper line). The unperturbed temperature profile is shown as the blue line and the perturbed temperature profile as the orange line. From left to right: Instantaneous RF: atmospheric temperatures are fixed everywhere; stratospheric-adjusted RF: allows stratospheric temperatures to adjust; zero-surface-temperature-change RF: allows atmospheric temperatures to adjust everywhere with surface temperatures fixed; and equilibrium climate response: allows the atmospheric and surface temperatures to adjust to reach equilibrium (no tropopause flux imbalance), giving a surface temperature change (ΔTs). From IPCC AR4 Chapter 2.



Climate response is the meteorological result of climate forcings, such as global temperature change, precipitation changes, or sea level changes.

Climates responses depend on the relative rate of changes in climate forcing versus response time of the climate system. Seasonal changes in tropical monsoons are an example of fast response to climate forcing, whereas ice sheets are an example of slow response.



Climate sensitivity, ΔT/ΔF, is the mean change in global temperature that occurs in response to a specified forcing.

The forcing will alter the balance energy budget:

$$\Delta F = \Delta Q_{TOA} - \Delta I_{TOA}$$

And to return to equilibrium the surface temperature of the climate systm changes by ΔT_s

$$\Delta T_s \left(\frac{\Delta I_{TOA}}{\Delta T_s} - \frac{\Delta Q_{TOA}}{\Delta T_s} \right) = -\Delta F = G$$

So that $\Delta T_s = \lambda G$ where λ is the climate sensitivity parameter which measures the climate response per unit forcing

$$\lambda = \left(\frac{\Delta I_{TOA}}{\Delta T_s} - \frac{\Delta Q_{TOA}}{\Delta T_s} \right)^{-1}$$

Climate sensitivity λ depends on many processes.
For longwave energy, links between T_s and total water vapor content M_{H2O} and cloud fraction A are very important:

$$\frac{\Delta I_{TOA}}{\Delta T_s} = \frac{\partial I_{TOA}}{\partial T_s} + \frac{\partial I_{TOA}}{\partial M_{H_2O}} \frac{\Delta M_{H_2O}}{\Delta T_s} + \frac{\partial I_{TOA}}{\partial A} \frac{\Delta A}{\Delta T_s} + \ldots$$

For shortwave energy, planetary albedo, cloud fraction, and vegetation are also important

$$\frac{\Delta Q_{TOA}}{\Delta T_s} = \frac{F_0}{4} \left(\frac{\partial R}{\partial T_s} + \frac{\partial R}{\partial A_c} \frac{\Delta A}{\Delta T_s} + \frac{\partial R}{\partial V} \frac{\Delta V}{\Delta T_s} + \ldots \right)$$

Considering T_{TOA} depends only on T_s, and $I_{TOA} \approx \varepsilon \sigma T_s^4$, the climate sensitivity is given by

$$\lambda = \left(\frac{\partial I_{TOA}}{\partial T_s} \right)^{-1} = (4 \varepsilon \sigma T_s^3)^{-1} = \left(\frac{4 \varepsilon \sigma T_s^4}{T_s} \right)^{-1} = \frac{T_s}{4 I_{TOA}}$$

For $T_s \approx 300, I_{TOA} \approx 240 W.m^{-2}, \lambda \approx 0.3 K (W.m^{-2})^{-1}$

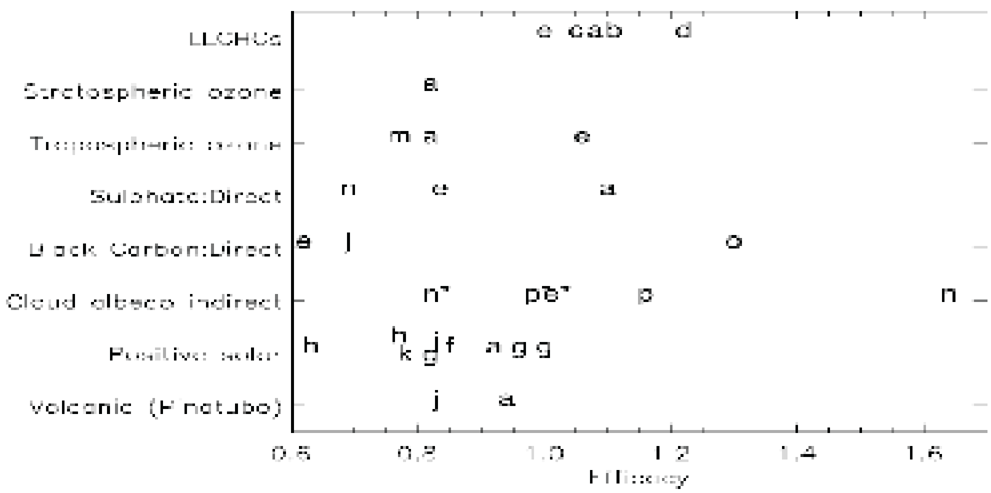
Climate sensitivity to doubling CO_2 : $0.5 \leq \lambda_{CO_2} \leq 1.5$ [K/(W.m^{-2})] which translates into a global warming of 2.0 K to 4.5 K for a doubling of CO_2 concentration

Efficacy and Effective Radiative Forcing

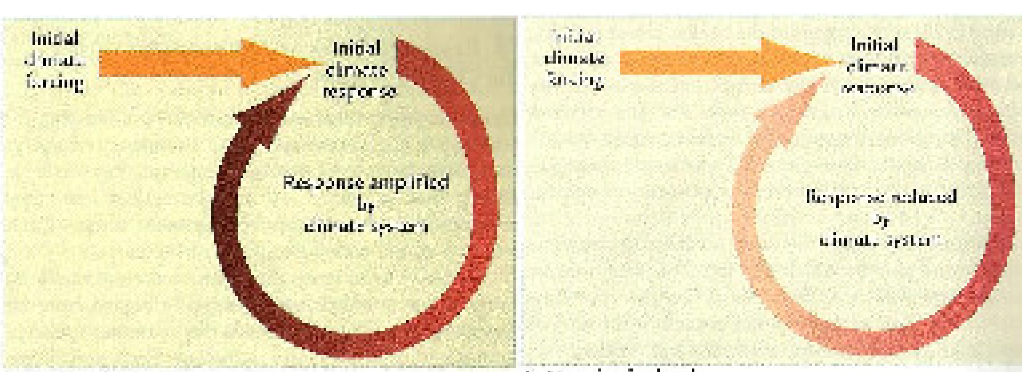
Efficacy is defined as the ratio of the climate sensitivity parameter for a given forcing agent (λ_i) to the climate sensitivity parameter for CO2 changes, $E_i = \dfrac{\lambda_i}{\lambda_{CO_2}}$

Efficacy can then be used to define an effective $RF = E_i\,RF_i$. For the effective RF, λ_i is independent of the mechanism, so comparing this forcing is equivalent to comparing the equilibrium global mean surface temperature change $\Delta T_s = \lambda_{CO_2} \times E_i \times RF_i$

The Figure below shows *Efficacies as calculated by several GCM for realistic changes of RF agents. Letters are centered on efficacy value and refer to the literature study that the value is taken from.* IPCC AR4 CH2.



Feedback is a relationship between two or more components of the climate system whereby changes in one component cause changes in the other(s), which in turn affect further change in the first component. This relationship may be positive or negative.



A. Positive feedback

B. Negative feedback

Examples of positive feedback: snow and ice have high albedo reflecting solar radiation away from the Earth's surface, thereby cooling the surface temperature, and, hence, the reduced temperatures tend to result in more snow and ice.

Negative feedback is a case when changes in the original component are damped out by its effect on the other components.