Today's Lecture (Lecture 11): Anthropogenic climate change — robust projections

Reference

IPCC AR5, Ch. 12
Holton, Ch. 11
Held and Soden (2006), Allen and Ingram (2002), Andrews et al. (2009), linked from course webpage

8 – Anthropogenic climate change

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8.1 – A brief history of anthropogenic climate change

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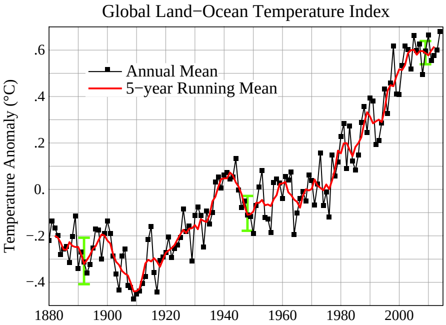
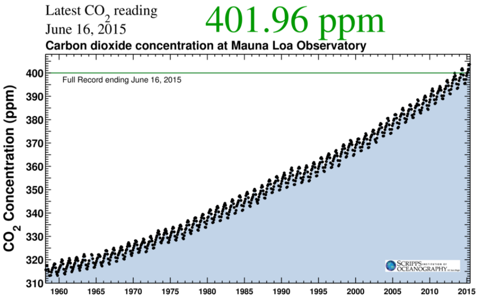
Some aspects of the instrumental record





Figures: GISS, SIO

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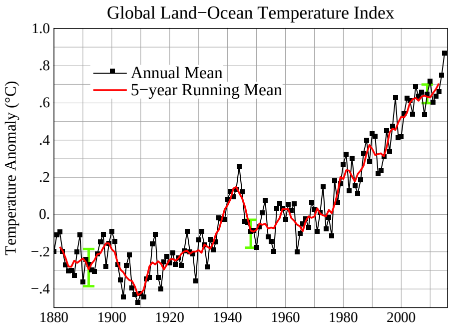

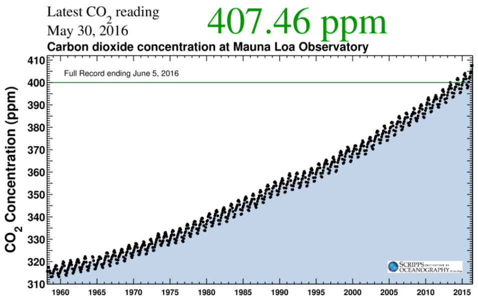

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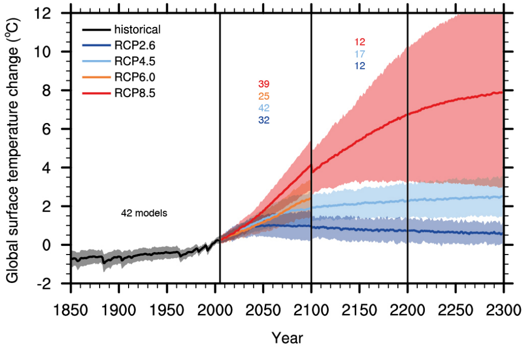


Projections are based on models – of emissions or concentrations of anthropogenic forcing agents (aerosols and greenhouse gases), possibly land use change, etc.

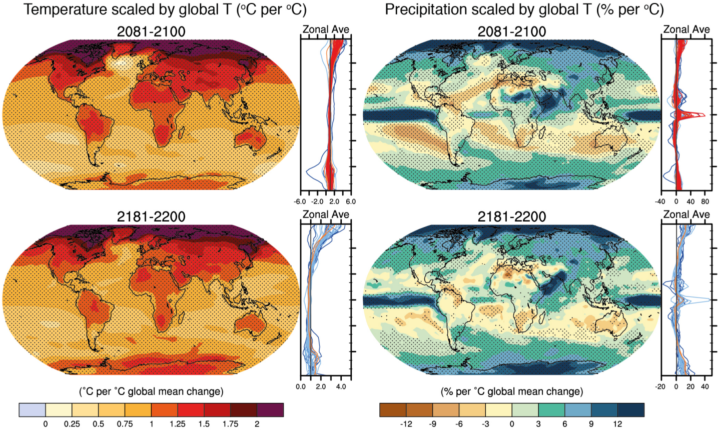
Figures: IPCC AR5 unless noted

Robust projections

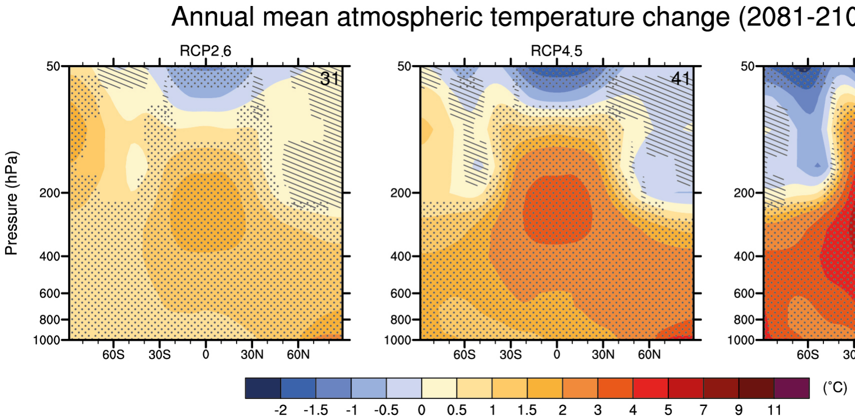
Robust projections are projections where models are in agreement and the underlying physical mechanisms are well understood. The main ones are

- increase in global-mean surface temperature
- arctic amplification
- precipitation pattern changes

Examples where models disagree: regional climate, magnitude of the warming. Clouds and aerosols are the biggest source of uncertainty, so they will get their own lecture.



Vertical–meridional structure of the warming



Moist adiabats become less steep with increasing temperature, recalling (2.59):

$$\Gamma_s = -\frac{\partial T}{\partial z} \approx \frac{g}{c_{pd} + L_{lv}(dq_s/dT)} = f(T) \neq \text{const}, \quad \frac{dq_s}{dT} \sim \exp(cT) \tag{8.1}$$

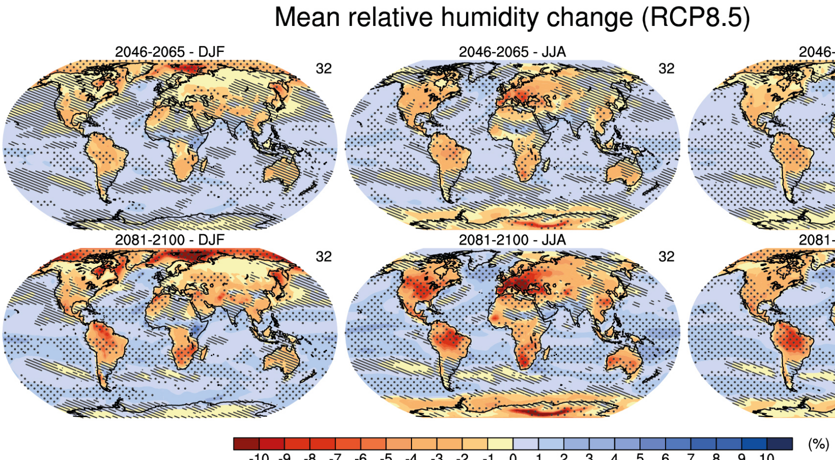
Γ_s is always smaller than the dry adiabatic lapse rate and decreases with increasing temperature.

Moisture scales with Clausius–Clapeyron equation

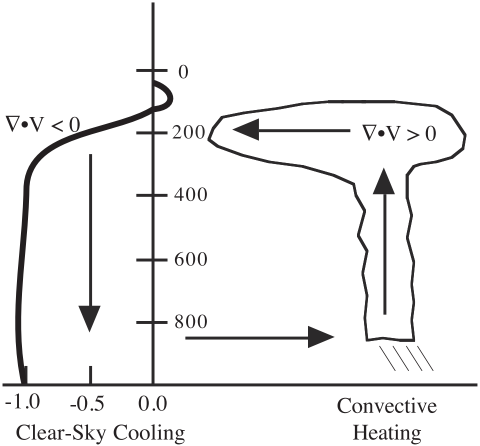
Moisture scales with Clausius–Clapeyron equation (CC, 2.54):

$$\frac{d \ln e_s}{dT} = \frac{L}{R_v T^2} \approx 7\% \text{ K}^{-1} \tag{8.2}$$

Evidence that relative humidity is relatively constant in a warming climate – processes that control relative humidity are not sensitive to climate change – so that $d \ln q/dT$ also approximately follows CC (7% K^{-1})



Balance between vertical motion, radiative cooling and convective heating



Updrafts: latent heating $Q \approx 10$ K day^{-1}, high ω

Subsidence: radiative cooling $Q \approx -1$ K day^{-1}, low ω

Steady-state ($\partial/\partial t = 0$), neglecting horizontal advection:

$$\omega \frac{T}{\theta}\frac{\partial\theta}{\partial p} = \frac{Q_{\text{rad}}}{c_p} \tag{8.3}$$

where

$$\sigma = \frac{T}{\theta}\frac{\partial\theta}{\partial p} \tag{8.4}$$

is called the *stability parameter*

Figure: Hartmann and Larson (2002)

Response of the tropical circulation to warming

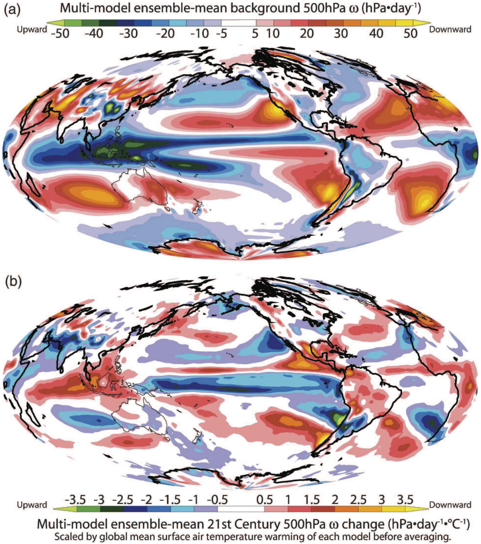


Figure: Vecchi and Soden (2007)

- Even in a warmer climate, the tropical atmosphere conserves θ_e:

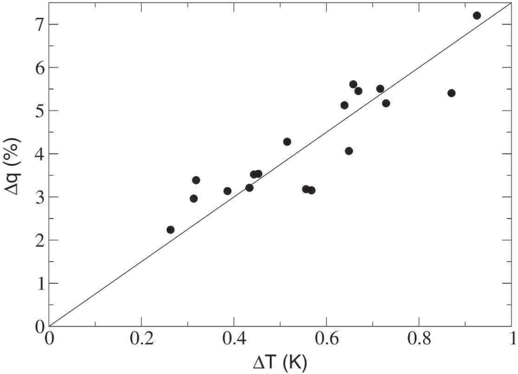
$$\theta_e = \theta \exp\left(\frac{L_{lv} q_s}{c_{pd} T}\right) \tag{8.5}$$

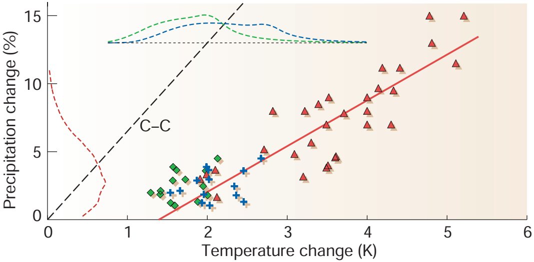
- In the upper troposphere, where $q_s \approx 0$, $\theta_e \approx \theta$
- In the boundary layer, where $q_s \sim$ CC, $\Delta\theta_e/\Delta\theta > 1$ with global warming
- Therefore $\Delta\theta(200\text{ hPa}) > \Delta T_S = \Delta\theta(1000\text{ hPa})$ and:
  - Average $\partial\theta/\partial p$ over the atmospheric column increases with Lq_s
  - Q_{rad} does not increase as fast (Knutson and Manabe, 1995)
  - Therefore

$$\omega = \frac{Q}{c_p}\left(\frac{T}{\theta}\frac{\partial\theta}{\partial p}\right)^{-1}$$

must decrease, resulting in a slowing circulation

Implications for precipitation





Think of the vertical circulation as air parcels leaving the boundary layer with high q, condensing and precipitating most of the BL q as they rise, and then returning to the surface with the low upper-tropospheric $q' \ll q$; so the precipitation is approximately $P = Mq$, where M is the vertical mass flux.

Since q scales approximately with CC in a warming climate (7% K^{-1}), but M decreases, the increase in precipitation is less than CC. (It turns out to be 2% K^{-1})

Figures: Held and Soden (2006), Allen and Ingram (2002)

Response of precipitation to warming — atmospheric energy budget

The precipitation response is evidently not constrained by water vapor availability (which increases according to CC). As we can see from the following argument, it is instead constrained by the atmospheric energy budget.

Recall that the atmosphere loses energy to radiative cooling at a rate

$$R_a = F_{\mathrm{TOA}} - F_s + R_{\mathrm{TOA}} - R_s \approx -100 \text{ W m}^{-2} \tag{8.6}$$

which is balanced by sensible and latent heat fluxes from the surface. Recall also that latent heat dominates.

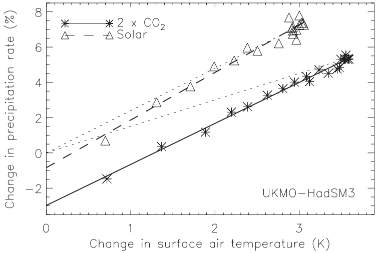
When the atmospheric energy fluxes are perturbed, equilibrium is restored on fast timescales due to the small heat capacity of the atmosphere. Thus, if we perturb the atmosphere by doubling the CO_2 concentration, it will quickly reequilibrate according to the equation

$$\Delta R_a = -(\Delta \mathrm{LH} + \Delta \mathrm{SH}) \approx -\Delta \mathrm{LH} = -L\Delta P \tag{8.7}$$

If we decompose the radiative perturbation into an ERF ΔR_c (forcing plus rapid adjustments) and a feedback ΔR_T (slow processes mediated by surface temperature increase), we can write the following equation for precipitation change:

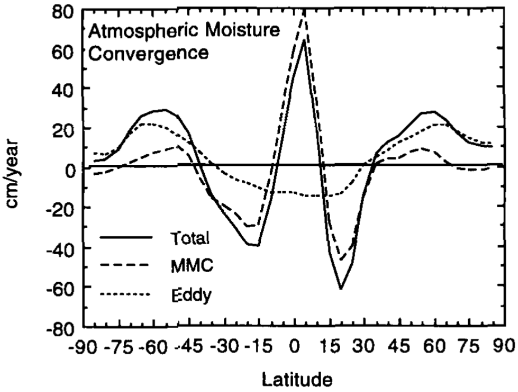
$$\begin{aligned} L\Delta P &= -\Delta R_c - \Delta R_T \\ &= -\Delta R_c + k_T \Delta T_S && (8.8) \\ &\approx -3 \text{ W m}^{-2} + 2 \text{ W m}^{-2}\text{ K}^{-1} \Delta T_S && (8.9) \end{aligned}$$

$$\frac{\Delta P}{P} \approx -3\% + 2\% \text{ K}^{-1} \Delta T_S \quad (\text{normalizing by } LP \approx 100 \text{ W m}^{-2}) \tag{8.10}$$



Allen and Ingram (2002); Andrews et al. (2009) (figure)

Geographic distribution of the precipitation response



- $P - E$, the difference between precipitation and evaporation, balances the convergence of atmospheric moisture transport:

$$P - E = \nabla \cdot \vec{F}, \quad \vec{F} = q\vec{v} \tag{8.11}$$

(this is the analog to convergence of atmospheric energy transport balancing the atmospheric energy budget). Consider F, the zonal-mean meridional component of $\vec{F}$.

- In a warming climate, meridional moisture transport changes because q increases (CC) and because the meridional circulation changes; CC dominates, so that

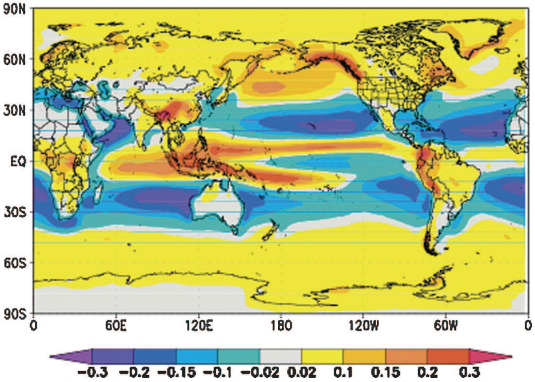
$$\frac{\Delta F}{F} \approx \frac{\Delta q}{q} = 0.07\ \mathrm{K}^{-1} \Delta T \tag{8.12}$$

- Therefore

$$\begin{aligned} \Delta(P - E) &= \nabla \cdot \Delta F \\ &\approx 0.07\ \mathrm{K}^{-1} \Delta T\, \nabla \cdot F \\ &\approx 0.07\ \mathrm{K}^{-1} (P - E) \Delta T \end{aligned} \tag{8.13}$$

Figure: Hartmann (1994)

Wet get wetter, dry get drier



- From (8.13), the change in the precipitation pattern under global warming is proportional to the precipitation pattern itself
- As a result, this pattern is called “wet get wetter, dry get drier”

Figure: Held and Soden (2006)