

# Extreme Precipitation

Nora Mascioli

# Extreme precipitation

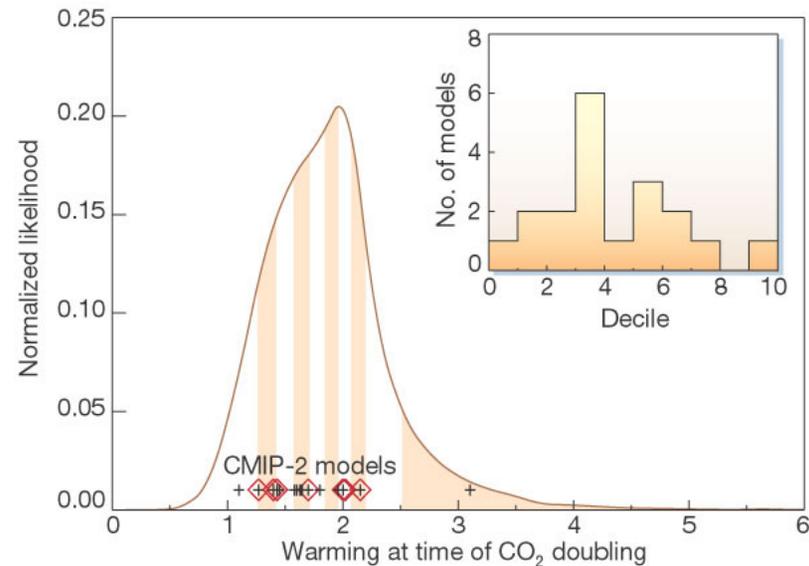
- How will precipitation change in response to anthropogenic climate change?
- What mechanisms drive changes in mean precipitation and changes in extreme precipitation?
- How can we use climate models to constrain the uncertainty in future precipitation?

# Global mean temperature change

- Global mean temperature response to increasing greenhouse gases is determined by:
  - Feedback strength (“climate sensitivity”)
  - The heat capacity of the upper ocean
  - How heat export to the deep ocean depends on recent surface changes
- These mechanisms are expected to change slowly if at all, and are often assumed to be constant.

# Transient Climate Response

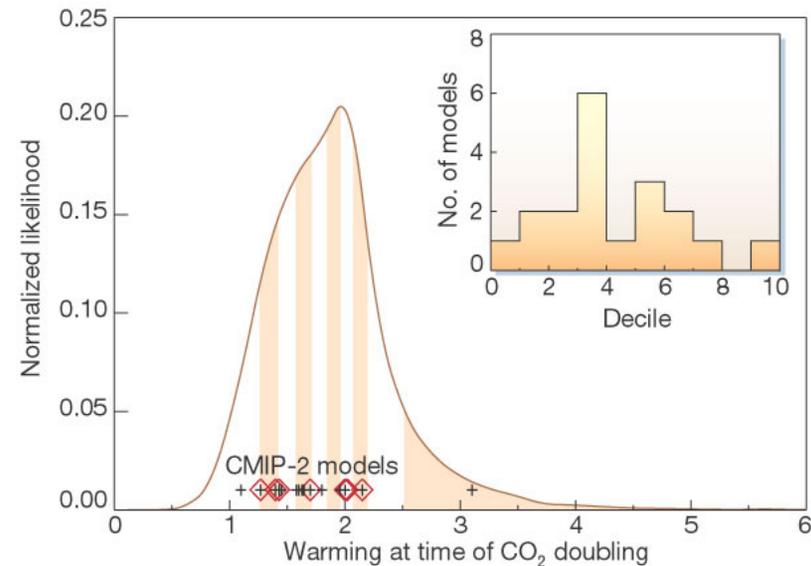
- We use observations of the near surface temperature changes and ocean heat content to constrain estimates of climate sensitivity.
- This allows us to create a probabilistic forecast of future changes in global mean temperature.



**Figure 1** Transient climate response (TCR), estimated from comparing an intermediate-complexity model with observations of recent large-scale temperature change. Vertical bands show equal-area deciles of the distribution. The crosses are the TCRs from CMIP2 models. Red diamonds show models used in the TAR summary range of '1.4–5.8 °C warming from 1990 to 2100'. The inset histogram shows how many of the CMIP2 models fall into each decile of the observationally constrained distribution.

# Transient Climate Response

- If climate models represented a random sample of possible climate responses consistent with observations, then we would expect the models to be evenly distributed between each decile.
- Each model represents a “best guess”, so model spread is not a good indicator of uncertainty.



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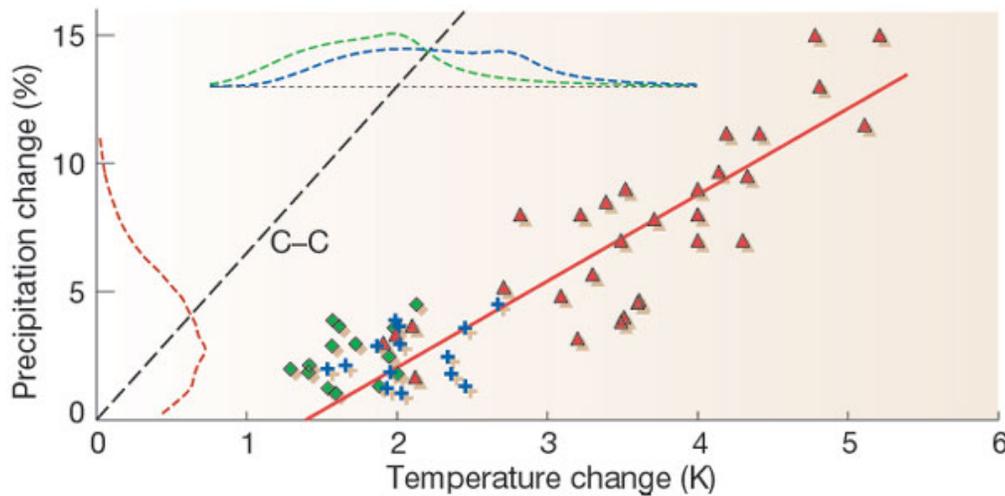
# Clausius-Clapeyron equation:

$$de_s/dT = L_v e_s / R_v T^2$$

- As T increases, the saturation vapor pressure  $e_s$  increases exponentially
- Relative humidity is expected to remain relatively constant, so the moisture content of the atmosphere is expected to increase

How does increased moisture availability in the atmosphere effect precipitation?

# The moisture budget



**Figure 2** Global mean temperature and precipitation changes in AOGCMs and corresponding probability distributions. Red triangles show global-mean T and precip changes in  $2\times\text{CO}_2$  slab ocean experiments. Green diamonds show the same, at the time of  $\text{CO}_2$  doubling, for those CMIP2 models for which the data are available. Blue crosses are the green diamonds adjusted for disequilibrium. The green dashed curve is the observationally constrained estimate of the distribution of global-mean temperature change at the time of  $\text{CO}_2$  doubling from Fig. 1. The blue curve is the same, but adjusted for disequilibrium. The red curve shows the distribution of global-mean precipitation changes implied by the blue curve, assuming the same relationship observed in the 'slab' experiments.

- If changes in precip are controlled by the C-C relationship, we would expect an increase of  $\sim 6.5\% \text{ K}^{-1}$  (black dashed line)
- Instead, models find a precip change of  $\sim 3.4\% \text{ K}^{-1}$  (red line)

Moisture availability is not controlling the overall intensity of the hydrological cycle in these models

# The energy constraint

- The overall intensity of the hydrological cycle is controlled by the availability of energy.
- The atmosphere must be able to radiate away latent heat released during precipitation.
- The tropospheric energy budget can be approximated by:

$$\Delta R_c + \Delta R_T = L\Delta P$$

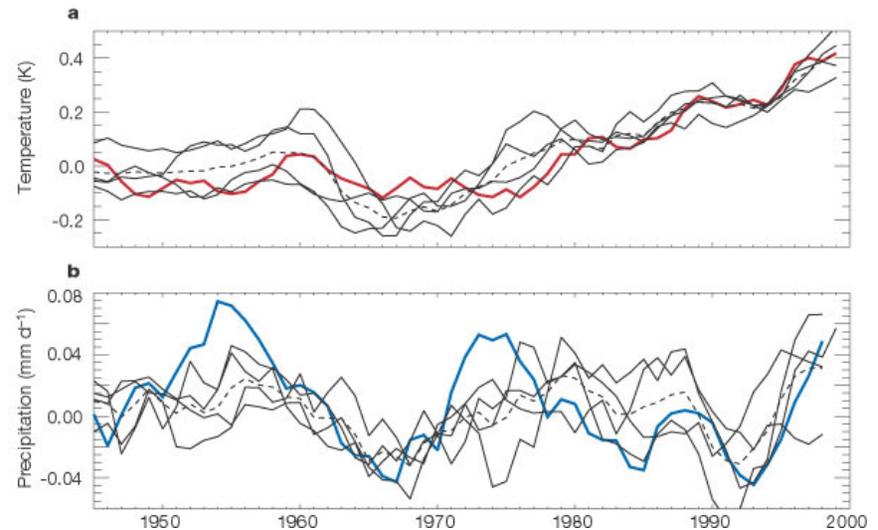
where  $L$  is the latent heat of evaporation,  $\Delta P$  is the change in global mean precip, and the perturbation radiative cooling,  $\Delta R$ , is split into a component,  $\Delta R_c$ , independent of  $\Delta T$ , and a component,  $\Delta R_T$ , dependent on  $\Delta T$ .

# How does CO<sub>2</sub> effect the energy budget?

- Doubling CO<sub>2</sub> decreases net outgoing IR radiation at the tropopause by 3-4 W m<sup>-2</sup> but increases downward IR flux at the surface by 1 W m<sup>-2</sup>, so  $\Delta R_c = -2$  or  $-3$  W m<sup>-2</sup>.
- So if tropospheric temperatures remained constant, CO<sub>2</sub> would decrease the intensity of the hydrological cycle.
- $\Delta R_T = k_T \Delta T$ , where  $k_T = 3$  W m<sup>-2</sup> K<sup>-1</sup>
- $\Delta T$  is estimated to be 1.8 - 6.5 K (Fig. 1), producing a rough estimate of  $\Delta P$  as 2 - 18%
- This is consistent with the estimate of  $\Delta P$  using the best fit line in Fig. 2 of 0.6 - 18%

# Observed changes in global (land) precipitation

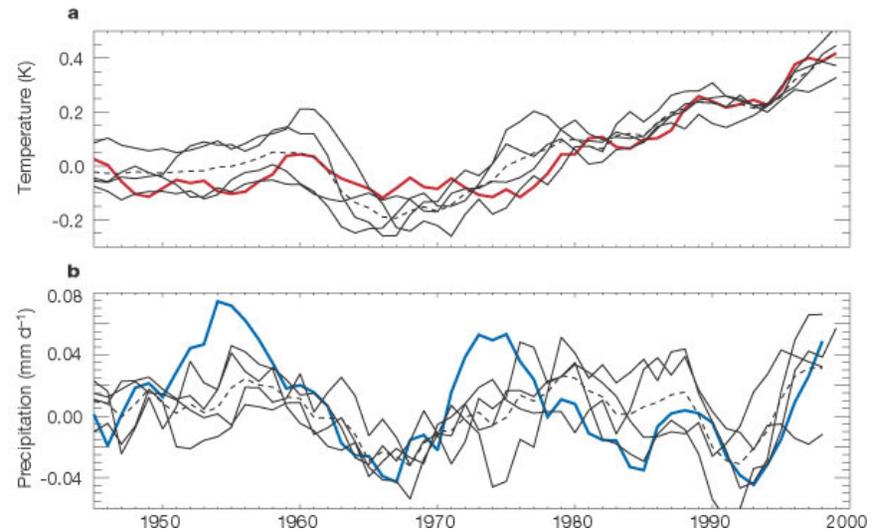
- $\Delta P/\Delta T$  depends on the nature of the forcing.
- For a change in shortwave radiation (such as aerosol forcing),  $\Delta R_c$  will be small, so  $\Delta P/\Delta T$  will be enhanced.
- The ensemble members and observations are correlated ( $r = 0.55$ ), indicating the influence of external forcing on global mean land precipitation.



**Figure 3** Change in global mean temperature (a) and land precipitation (b). The model data are from a 4-member ensemble of the HadCM3 climate model run with estimates of anthropogenic, solar and volcanic forcing (dashed line shows ensemble mean). Model data were included only where and when observations exist and a five-year running mean was applied.

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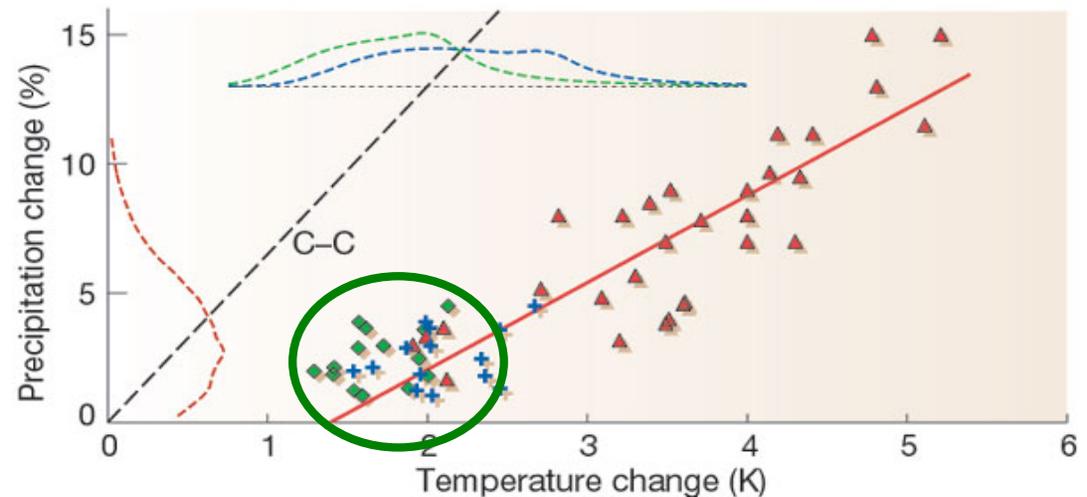
- However, it is clear that  $\Delta P$  is not simply following  $\Delta T$ .
- We cannot use the historical record to estimate  $\Delta P/\Delta T$ , because we cannot expect the relationship to remain constant.
- The anthropogenic contribution to  $\Delta P$  is still too small relative to natural external forcings and internal variability.



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# Ocean heat fluxes

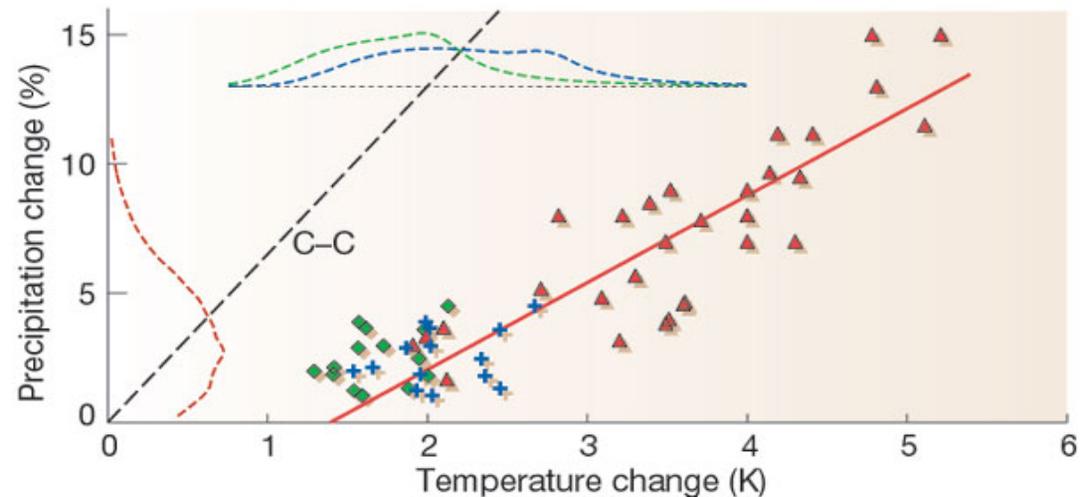
- The CMIP2 transient responses lie slightly above the best fit equilibrium line.
- We need to add a term to allow for heat flux into the oceans,  $F_s$ :



$$\Delta R_c + k_T \Delta T - \kappa F_s = L \Delta P$$

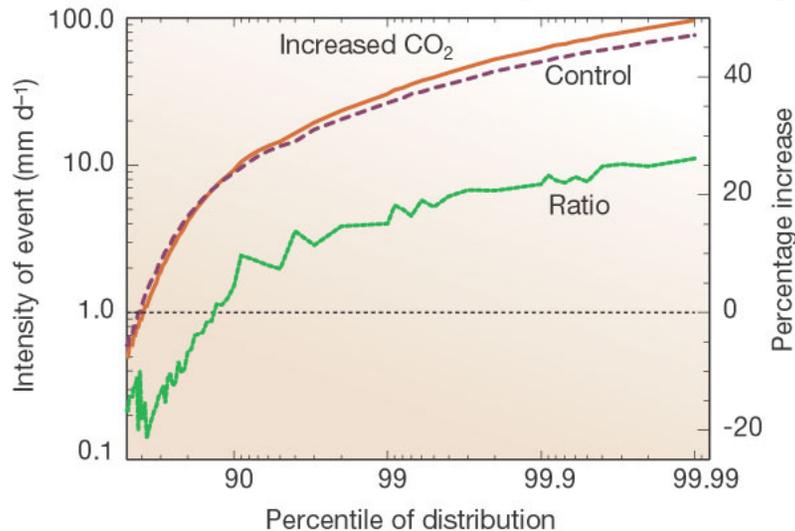
# Ocean heat fluxes

- The blue dashed line represents  $k_T \Delta T - \kappa F_s$  (modified from the transient climate simulations).
- Using the best fit line, Allen and Ingram construct a probability distribution for  $\Delta P$  under transient climate simulations (red dashed line).



The key takeaway: the uncertainty in  $\Delta P$  is much larger than the model spread.

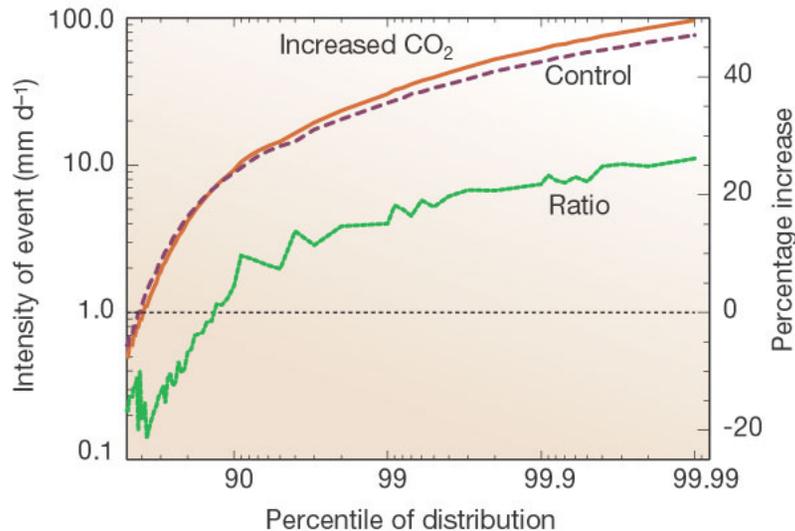
# What about extreme precipitation?



**Figure 4** Distribution of daily precipitation (all locations and seasons included) in years 2070–2100 of a transient climate-change simulation (red) and in the corresponding control simulation (purple). Their ratio is shown by the green curve and right-hand (linear) axis. The global-mean warming is 3.6 K and the tropical mean 3.3 K, giving a Clausius–Clapeyron limit on this ratio of about 22%.

- Heavy rainfall events occur when all water in a volume of air is precipitated out.
- The intensity of heavy rainfall events is determined by moisture availability.
- We expect extreme rainfall events to increase by  $6.5\% \text{ K}^{-1}$  (from the C-C relation) or more

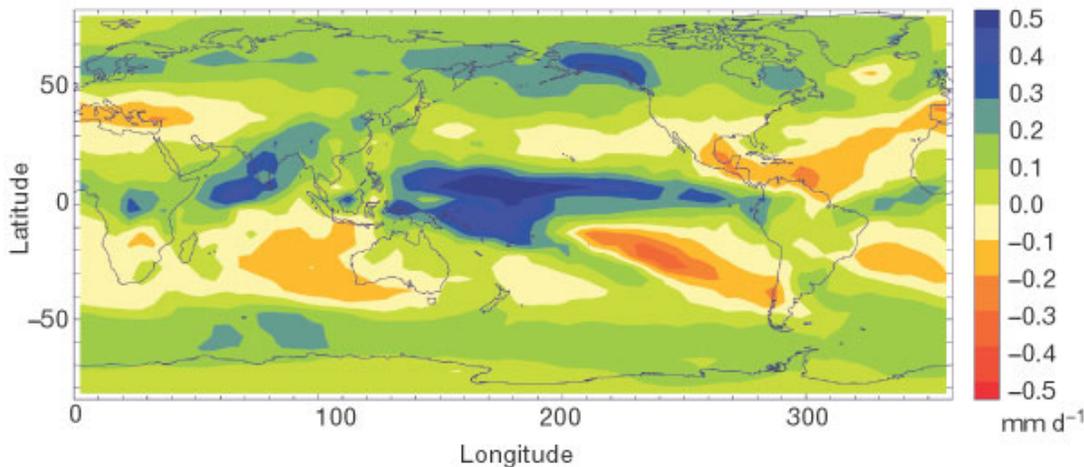
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- Figure 4 shows the difference in intensity of extreme rainfall events between the transient climate simulation and control run of a GCM.
- The most extreme events occur in the tropics, where the ratio is converging to ~25%, slightly higher than the value predicted by the C-C equation.
- These increases are much larger than the global mean increase.
- Rainfall must be decreasing elsewhere in the distribution to compensate.

# Regional precipitation

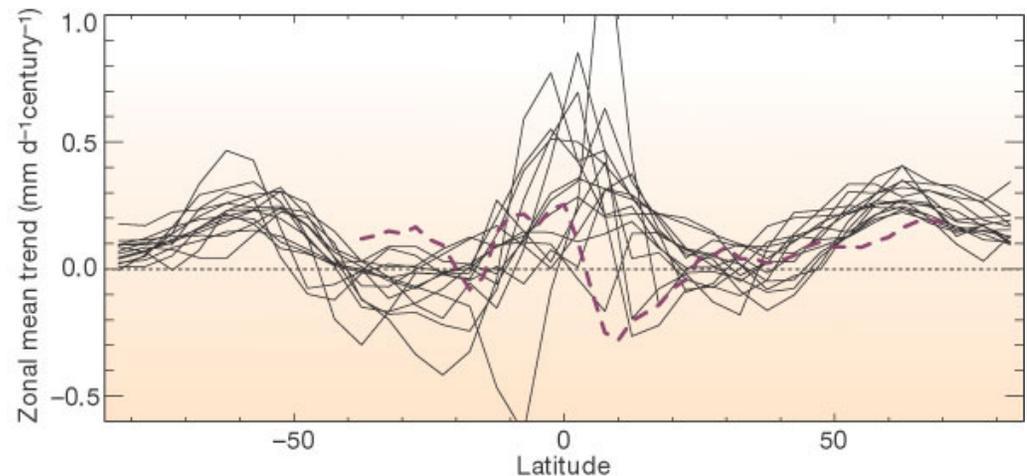


**Figure 5** Multimodel mean precipitation change at the time of CO<sub>2</sub> doubling (CMIP2), for the 20 years centered around the time of CO<sub>2</sub> doubling.

- Only 30% of the variance between the models is predicted by the mean response.
- Between 50S and 50N, the standard deviation of the models is larger than the mean.
- However, there may still be common patterns in the model response.

# Zonal mean precipitation

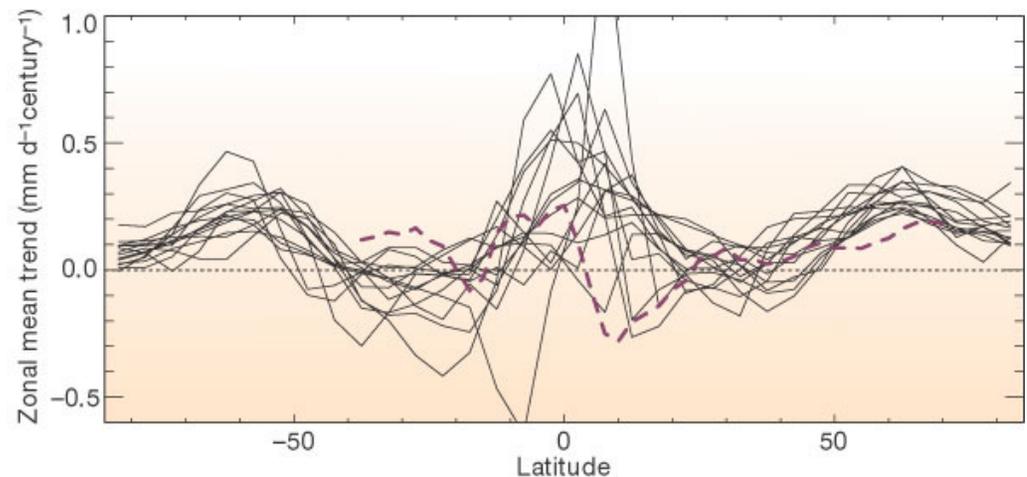
- There is less inter-model variability in the zonal mean.
- Models predict large increases in the tropics, slight decreases in the subtropics, and increases in the high latitudes.
- Models show mid-continental summer droughts intensifying as a result.



**Figure 6** Zonal-mean precipitation trends over the 70 years to CO<sub>2</sub> doubling in the CMIP-2 ensemble (solid) and observed (dashed) trends in terrestrial precipitation over the past century.

# Zonal mean precipitation

- Comparisons with the observed 100-year precipitation trends show some similarities.
- However, the observations are land-based, and the forcing over the historical period is very different from the forcing in the transient climate simulations.

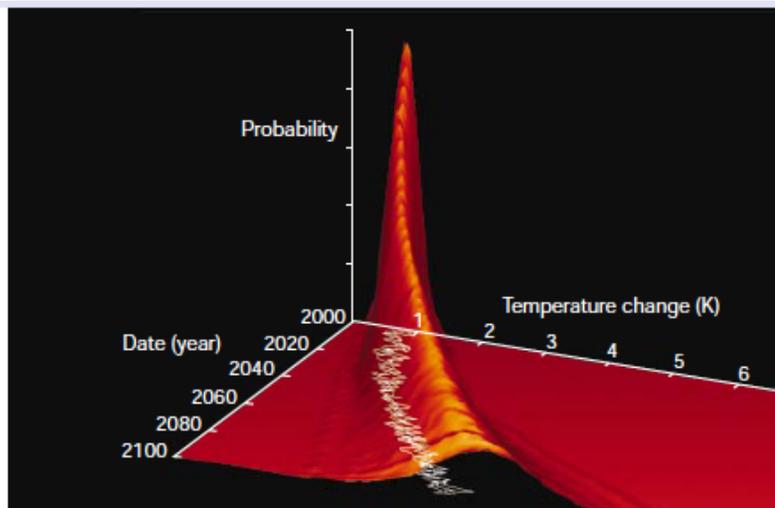


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# Aerosols and Oceans

- Aerosols (anthropogenic and natural) have a large effect on mean precipitation. Sulfate aerosols in particular reduce the incoming SW radiation, implying a small  $\Delta R_c$  and large  $L\Delta P$ .
- Aerosols can also heat the atmosphere, producing  $\text{CO}_2$  -like effects on the hydrological cycle.
- Aerosols could potentially change the sensible heat flux between the surface and the troposphere.
- However, uncertainties related to aerosol forcings are much larger than those related to well-mixed greenhouse gases.
- Increases in high latitude rainfall could weaken the thermohaline circulation. Most climate models show some weakening of the THC.

# Probabilistic forecasts



**Box 1 Figure** The three-dimensional surface shows a forecast probability distribution of a one-dimensional quantity (global-mean warming above pre-industrial), accounting for uncertainty in the climate response, while the lines show the (smaller) impact of initial condition uncertainty in an ensemble of model simulations. Data courtesy of P. Stott (Met Office) and J. Kettleborough (Rutherford Appleton Laboratory)<sup>5</sup>, based on the IPCC SRES A2 scenario.

- Because climate models represent a “best guess” of the climate system, intermodel spread is not a good measure of the uncertainty in the system. This makes probabilistic climate forecasting difficult.
- In order to perform a probabilistic forecast, we need a very large ensemble experiment obtained by systematically varying key model parameters, resolution, etc.

# Physical drivers of changes in precipitation extremes

- We expect changes in extreme precipitation to scale with the availability of moisture in the atmosphere.
- However, idealized model simulations show that the rate of increase of extreme precipitation varies by latitude:
  - Outside of the subtropics, the extremes vary similarly to the mean
  - In the tropics, there is large intermodel spread, however observations suggest are more sensitive to temperature change than expected.

# Observations vs CMIP3

- In general, the models are comparable to the observations, however the observations are very uncertain.
- Model spread is very large in the tropics
- Precipitation extremes increase at all latitudes, with maximum increases in the tropics.

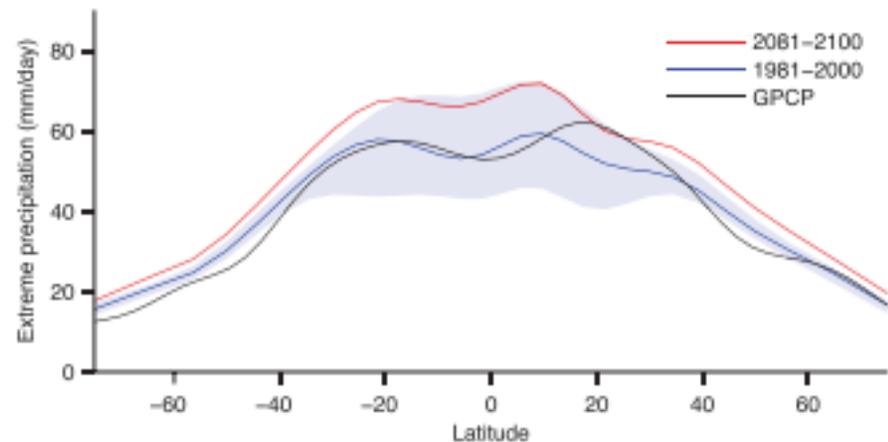


Fig. 1. The 99.9th percentile of daily precipitation (millimeters per day) for the periods 1981–2000 (blue) and 2081–2100 (red) in the SRES A1B scenario (multimodel median), and based on Global Precipitation Climatology Project (GPCP) data for the period 1997–2006 (black). Model scatter (shading) for the period 1981–2000 is shown using the interquartile range (50% of models lie within the shaded region). The spatial resolution of the GPCP data were degraded from 1° to 3°, which is comparable with climate model resolutions. A Gaussian smoothing filter of standard deviation 6° latitude was applied to reduce noise in all plots showing variations with latitude.

# Extreme precip vs water vapor content

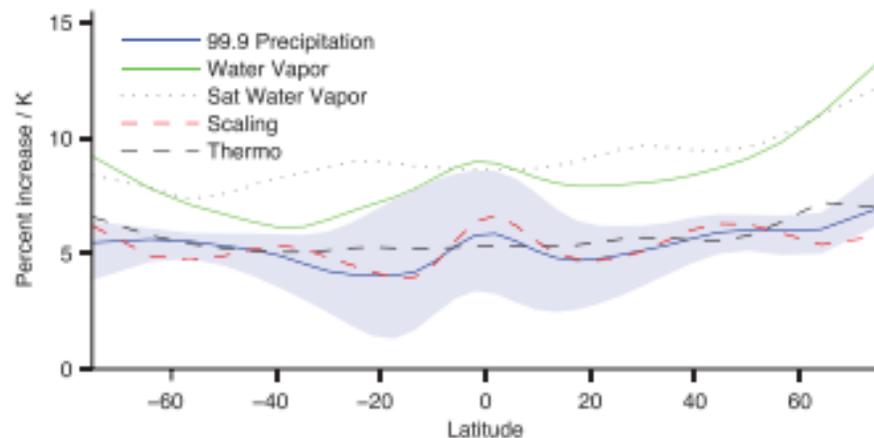
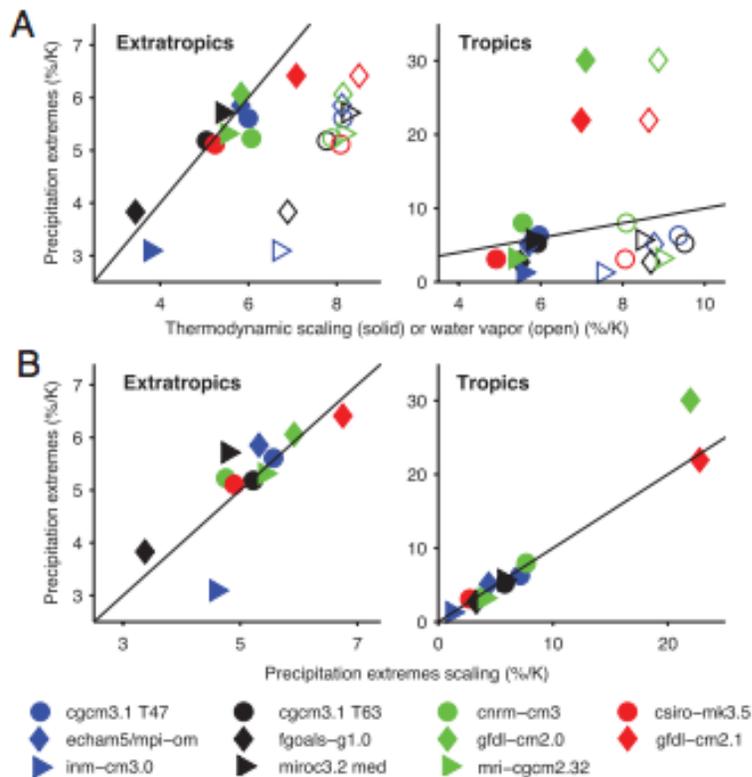


Fig. 2. Fractional changes in the 99.9th percentile of daily precipitation (blue), zonally averaged atmospheric water vapor content (green), saturation water vapor content of the troposphere (black dotted), full precipitation extremes scaling (Eq. 2) (red dashed), and thermodynamic scaling for precipitation extremes (black dashed). The lines show multimodel medians of the fractional changes relative to 20th-century values, normalized by the global-mean change in surface air temperature for each model. Model scatter is shown for the fractional change in precipitation extremes using the interquartile range (shading). The saturation water vapor content is calculated using an average of the climatological monthly-mean temperature over all times and longitudes at which the extreme precipitation occurs.

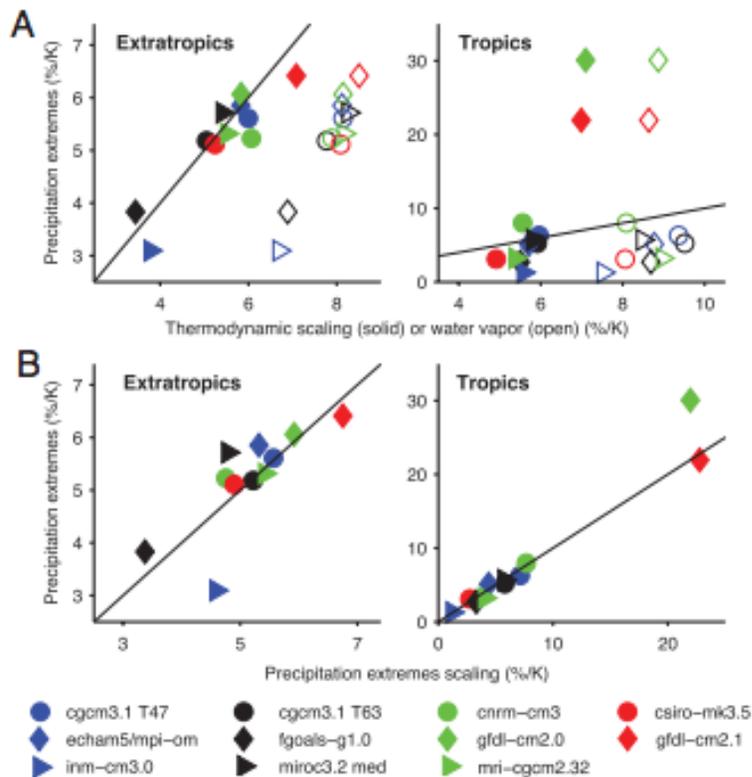
- Both extreme precip and water vapor content increase at all latitudes.
- However, the increases in extremes are consistently less than the increases in water vapor content.
- The central finding is the same if we consider saturation water vapor instead of mean water vapor.

# Comparing drivers of extreme precip



**Figure 3** Fractional changes in the 99.9th percentile of daily precipitation for each model versus changes in atmospheric water vapor content and scalings for precipitation extremes. (A) Atmospheric water vapor content (open symbols) and the thermodynamic scaling that neglects changes in upward velocity (closed symbols). (B) Full scaling for precipitation extremes. The fractional changes are relative to 20th-century values.

# Comparing drivers of extreme precip



- Extratropical precipitation extremes increase less rapidly than water vapor content in all models
- Tropical precipitation extremes increase at the same rate or less rapidly than water vapor content in most models (the GFDL models are the outliers)

# Precipitation extremes scaling

- During a precipitation event:
  - Air rises and cools adiabatically
  - Water vapor condenses
  - Latent heat is released
  - Condensate precipitates
- The condensation rate needed to maintain saturation levels in the air parcel depends on the vertical velocity,  $\omega$ , and the derivative of the saturation specific humidity,  $q_s$ :

$$c = -\omega \left. \frac{dq_s}{dp} \right|_{\theta^*}$$

# Static stability

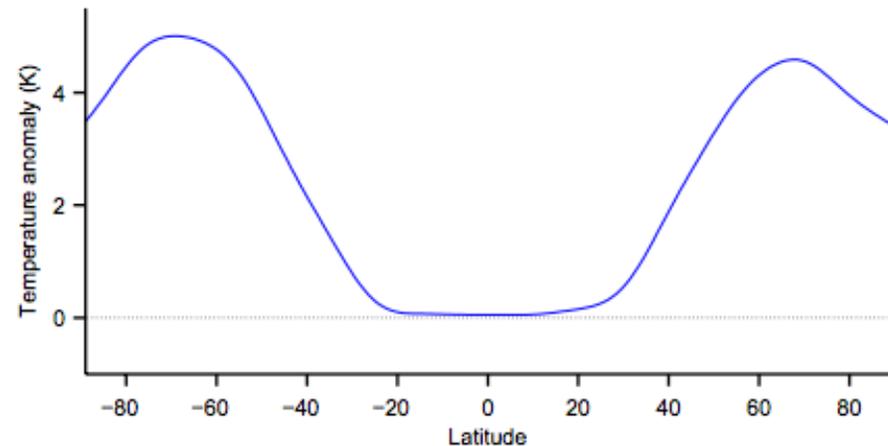
- This equation can be written in terms of the latent heat release:

$$c = \omega c_p T / (L \theta) d\theta / dp |_{\theta^*}$$

- The latent heat release from condensation balances the adiabatic cooling in the updraft.
- The static stability changes more slowly with temperature than the saturation specific humidity, resulting in the relatively slow increase in precipitation extremes

# Precipitation and Temperature

- Extreme precipitation is highly temperature dependent.
- 600hPa extratropical temperatures tend to be up to 5K higher than the climatological mean on days with heavy precipitation.



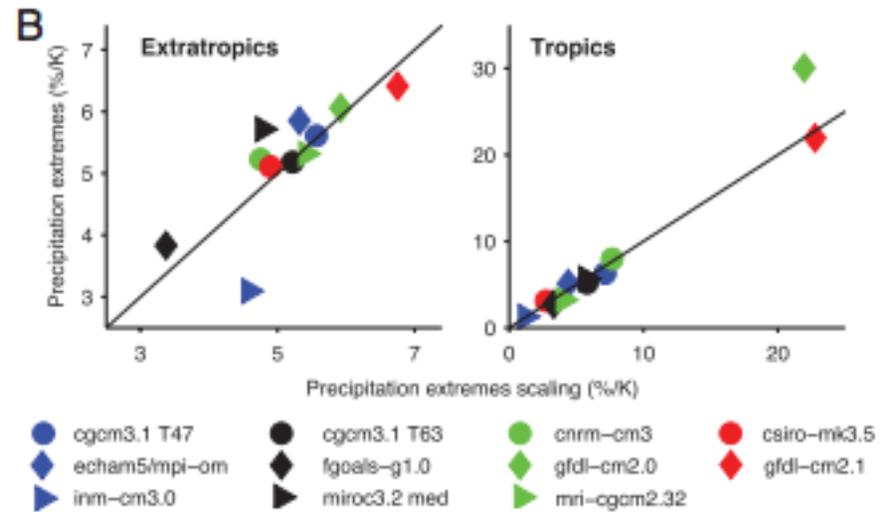
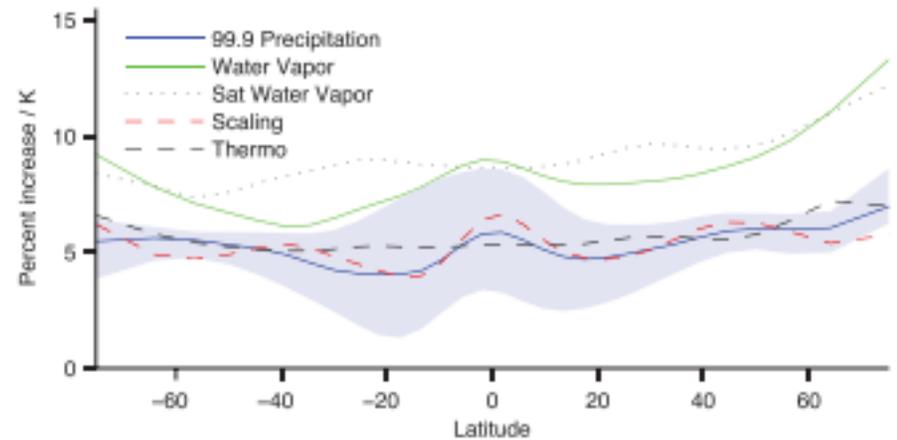
**Figure S3** 600hPa Temperature anomalies associated with 99.9th percentile precipitation events

# Precipitation intensity

- Extreme precipitation,  $P_e$ , can be expressed at each latitude as:

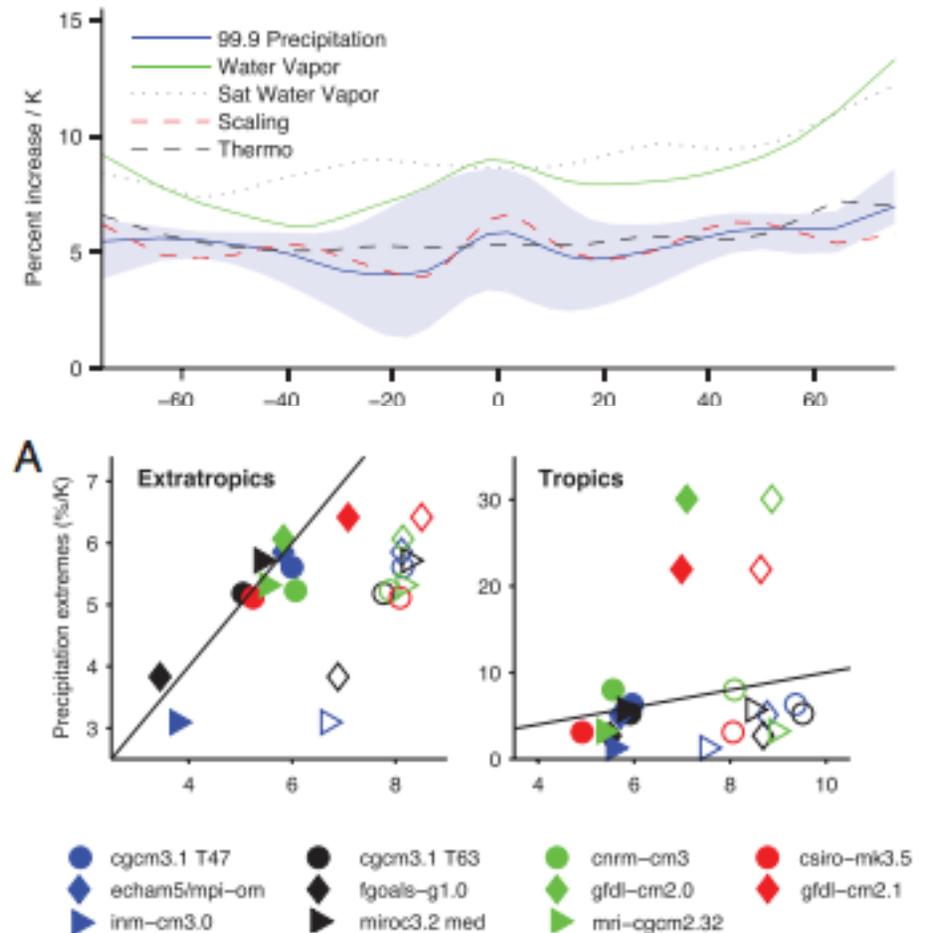
$$P_e \sim - \left\{ \omega_e \frac{dq_s}{dp} \Big|_{\theta^*, T_e} \right\}$$

- This scaling captures the behavior of precipitation extremes for all latitudes, in all models except for one outlier (GFDL)



# Thermodynamic contribution

- By removing  $\omega_e$ , O’Gorman and Schneider evaluate the relative contributions of the thermodynamic component and the updraft velocities.
- Fig. 2 and 3a show that this produces a good fit for precipitation extremes in the extratropics, but not in the tropics.
- Intermodel variability in extreme precipitation is driven by intermodel variability in  $\omega_e$ .



# Conclusions

- Intermodel spread between climate models underestimates the uncertainty in changes in precipitation.
- Mean precipitation depends on available energy, rather than available water.
- Extreme precipitation is theoretically dependent on available water.
- However, in practice, extreme precipitation increases less rapidly than moisture availability.
- Extreme precipitation can be scaled to changes in temperature and vertical velocity.
- Models are currently unable to predict changes in extreme precipitation in the tropics due to their wide spread of updraft velocities.