The dynamics of precipitation extremes

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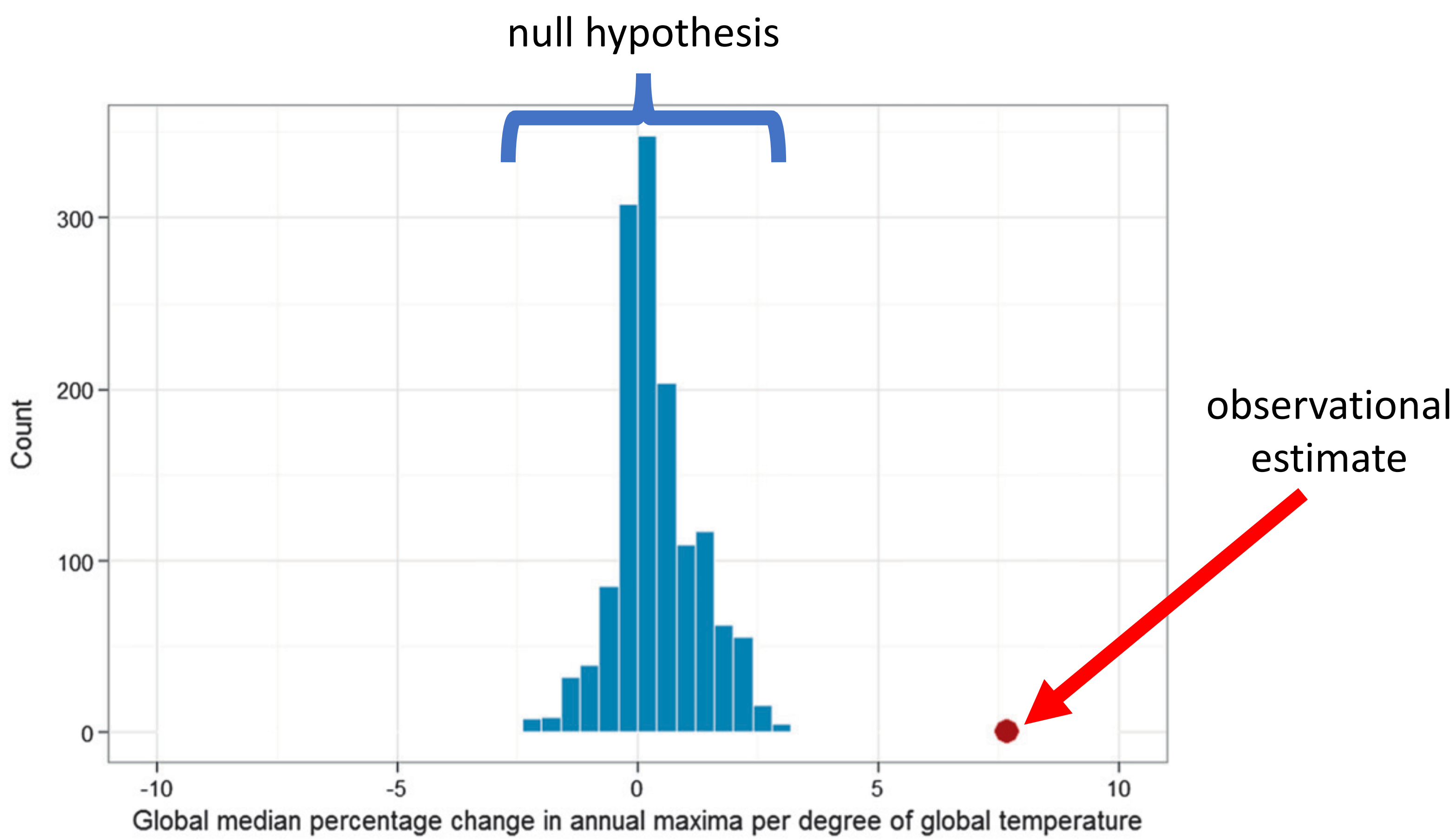
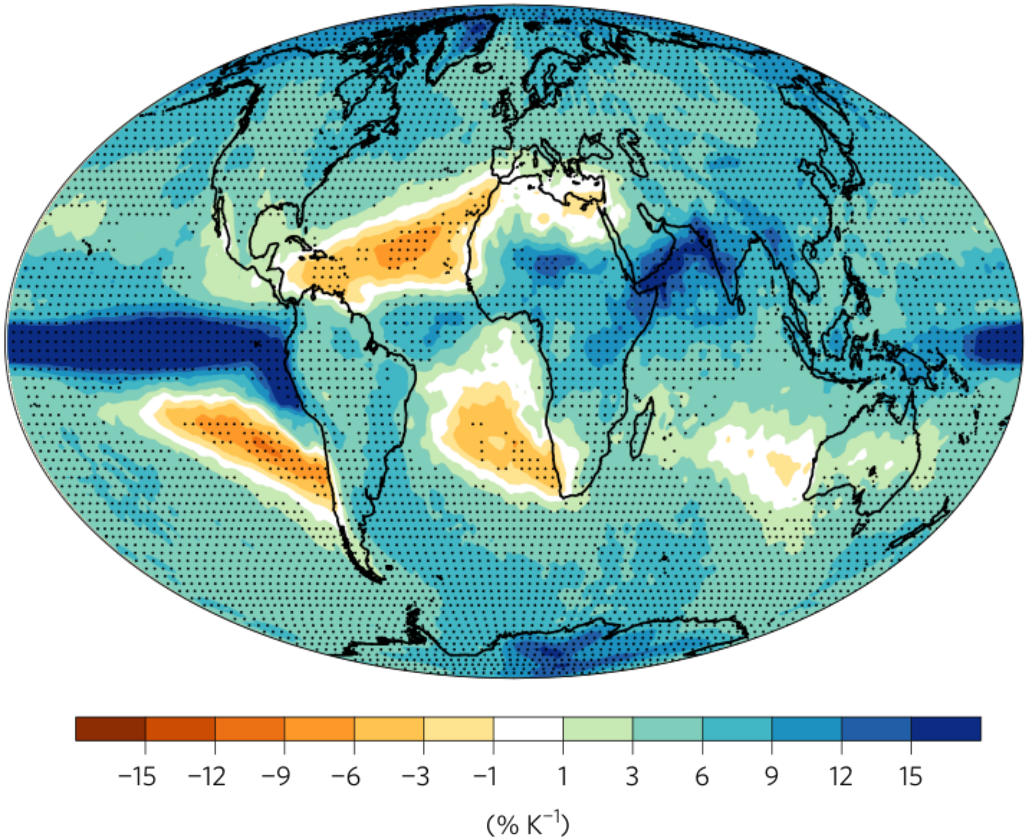
Observations indicate higher global temperature associated with more intense precipitation extremes



Observational estimate of sensitivity of precipitation extremes to global-mean temperature
(From Westra et al., 2013)

Climate model simulations also predict future increases in the intensity of precipitation extremes

Change in annual-maximum daily precipitation per unit global warming



Multi-model mean across 22 CMIP5 models under the RCP8.5 scenario. Stipling where 80% of models agree on sign of change
(From Pfahl et al., 2017)

- Why do we expect precipitation extremes to increase with warming?

- What sets the magnitude and spatial pattern of changes in precipitation extremes?

- Why do models disagree on the magnitude of future changes in precipitation extremes?

What sets the precipitation rate?

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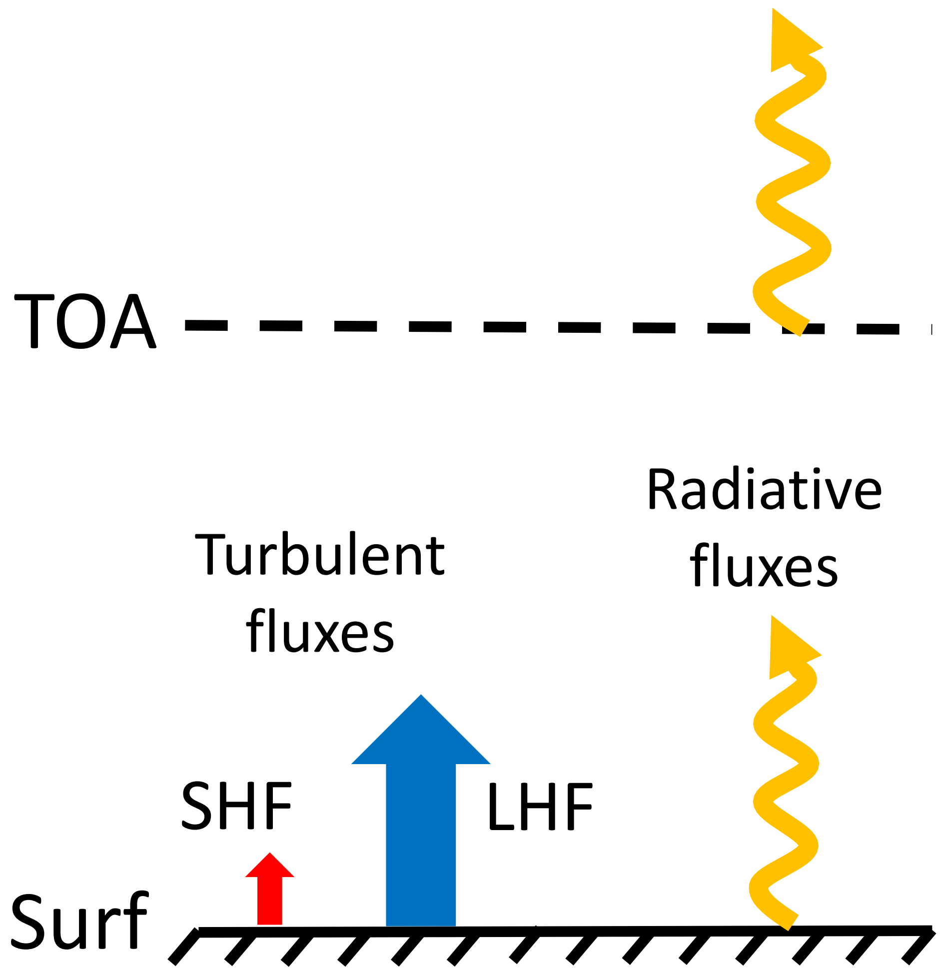
Global-mean precipitation

What sets the precipitation rate?

Global-mean precipitation

- The global-mean precipitation rate is set by the energy budget of the atmosphere
- This increases at roughly 2-3% per Kelvin global warming (Allen & Ingram, 2003)



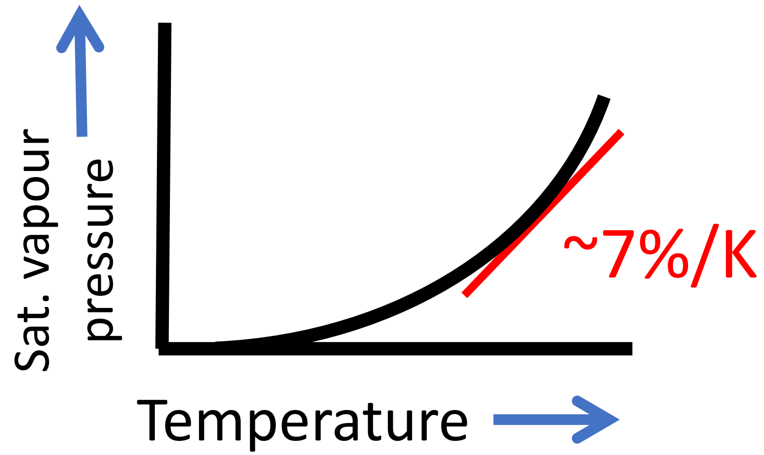
What sets the precipitation rate?

What about precipitation extremes?

What sets the precipitation rate?

What about precipitation extremes?

- Heaviest rainfall constrained by available water vapour? (Trenberth, 1999)



Can we make this a bit more quantitative?

A theory for precipitation extremes

Consider a large-scale precipitation extreme event.

For saturated air, the condensation rate may be approximated

$$c \approx -\frac{Dq_s}{Dt}. \tag{1}$$

Note that q_s is a thermodynamic function, and it may be written,

$$q_s = q_s(p, \theta_e^*),$$

where θ_e^* is the saturation equivalent potential temperature

A theory for precipitation extremes

We assume that, during an extreme event, the latent heating term dominates the thermodynamic equation so that,

$$\frac{D\theta_e^*}{Dt} \approx 0. \quad (2)$$

This allows us to write the condensation rate

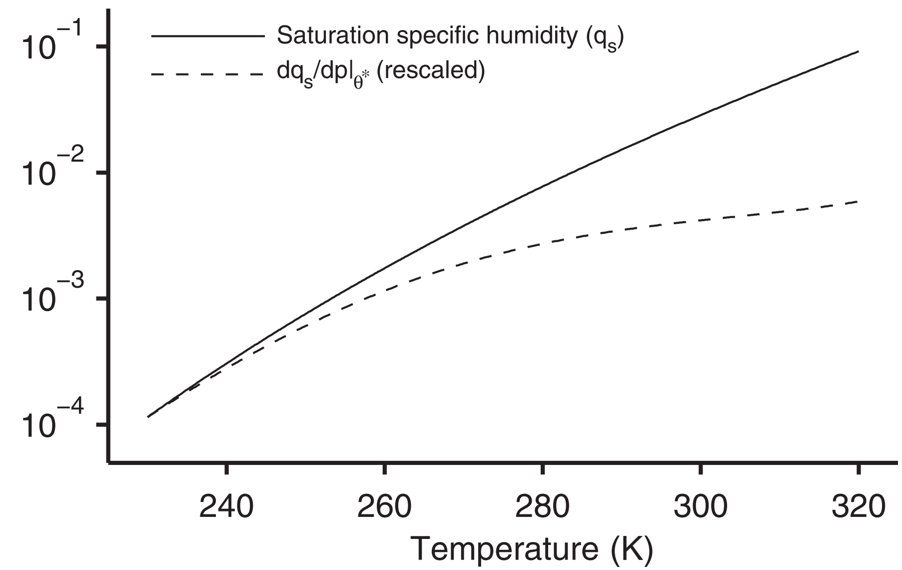
$$c = -\frac{Dp}{Dt}\frac{dq_s}{dp}\bigg|_{\theta_e^*} = -\omega\frac{dq_s}{dp}\bigg|_{\theta_e^*}. \quad (3)$$

Derivatives along a moist adiabat

The derivative with respect to saturation humidity is taken along a moist adiabat:

$$\left.\frac{dq_s}{dp}\right|_{\theta_e^*}.$$

It increases with temperature, but not as rapidly as the saturation humidity itself.



From O'Gorman & Schneider, 2009

A theory for precipitation extremes

Now, we relate the precipitation rate P to the vertically integrated condensation rate by an efficiency ϵ so that,

$$P = -\epsilon \int \omega \left.\frac{dq_s}{dp}\right|_{\theta_e^*} \frac{dp}{g}. \qquad (4)$$

where all quantities are evaluated at the time of the extreme event

Note: the efficiency accounts for the microphysical processes that convert condensation into precipitation, as well as the approximations used to estimate the condensation rate.

A theory for precipitation extremes

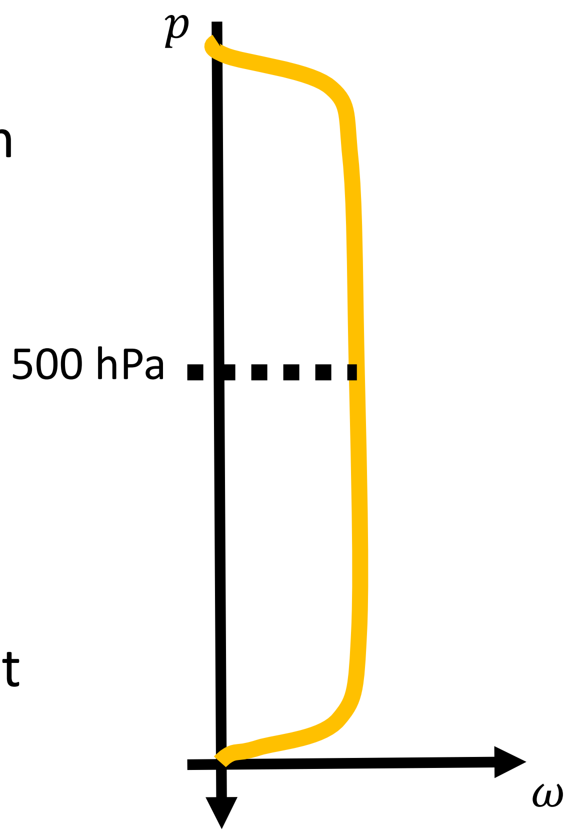
Assume:

1) the vertical velocity profile is constant apart from near the surface and at the tropopause

2) The atmospheric thermal structure is roughly moist adiabatic

Then the scaling equation may be integrated so that

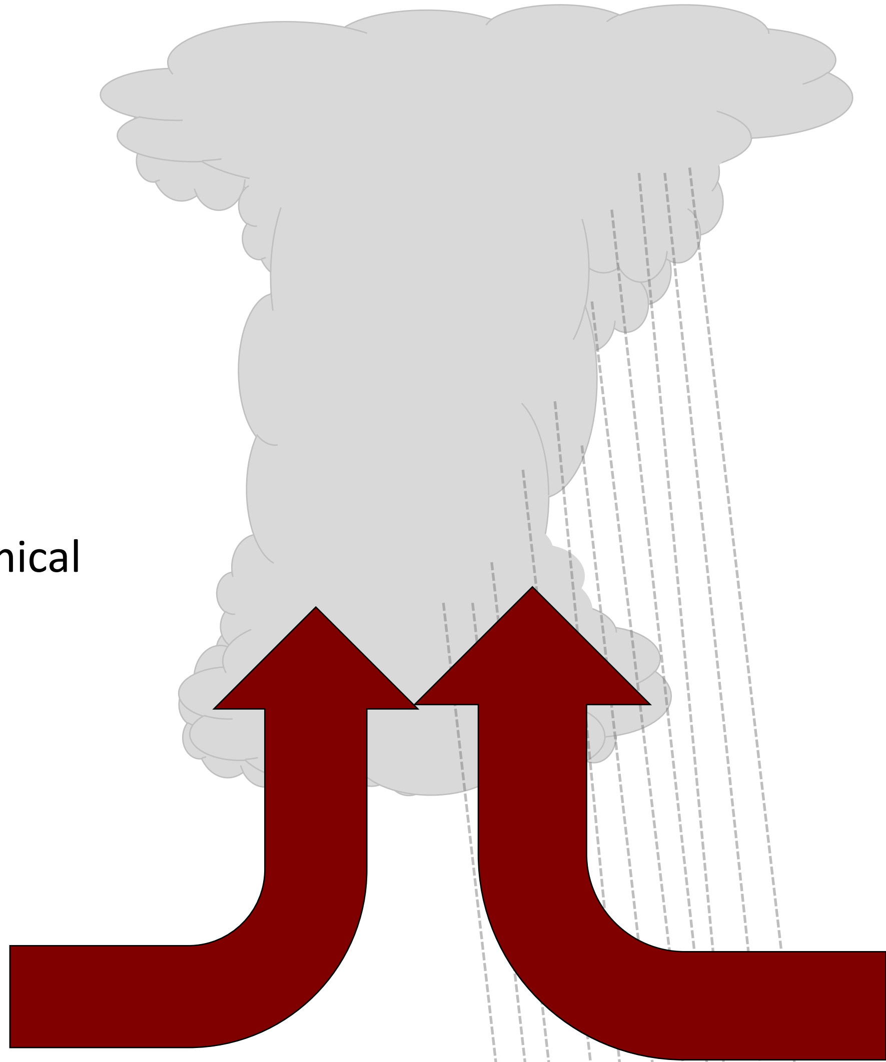
$$P \propto \epsilon \omega_{500} q_s(T_{BL}, p_{BL}).$$



Scaling of precipitation extremes with warming

$$\frac{\delta P}{P} \approx \frac{\delta \epsilon}{\epsilon} + \frac{\delta \omega_{500}}{\omega_{500}} + \frac{\delta q_s}{q_s}.$$

microphysical dynamical thermodynamical



Simple thermodynamic scaling for precipitation extremes

- If the efficiency and vertical velocity remain constant, precipitation extremes scale with the near-surface saturation specific humidity

- Increases at roughly 7%/K near the surface.

- This is often referred to as Clausius-Clapeyron scaling.

- The approximations used to derive this are not well satisfied, particularly outside the tropics

A more accurate scaling for precipitation extremes

More generally, we can use the theoretical equation (4):

$$P = -\epsilon \int \omega \frac{dq_s}{dp}\bigg|_{\theta_e^*} \frac{dp}{g} \qquad (4)$$

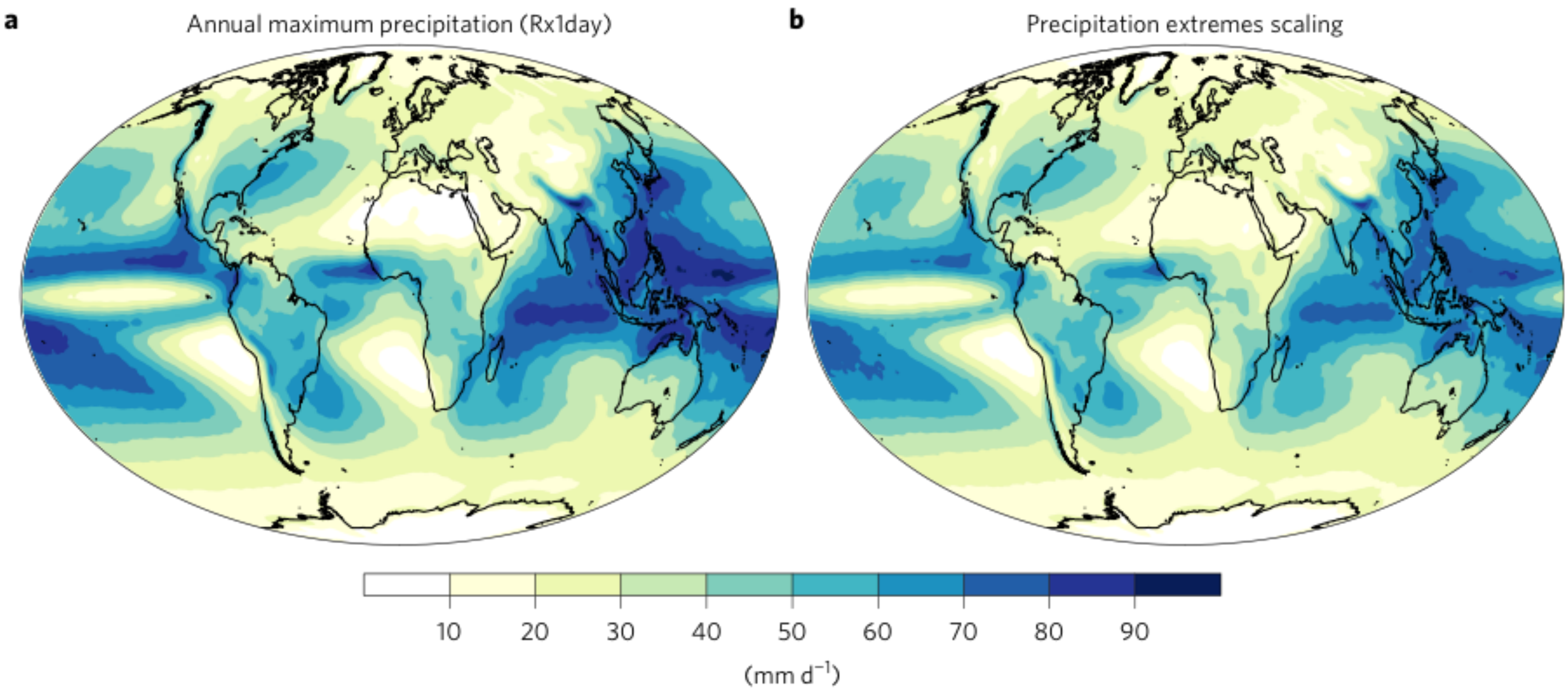
to understand changes to precipitation extremes

In particular, we assume the efficiency is independent of time and space to construct a scaling:

$$P \propto \int \omega \frac{dq_s}{dp}\bigg|_{\theta_e^*} \frac{dp}{g}. \qquad (5)$$

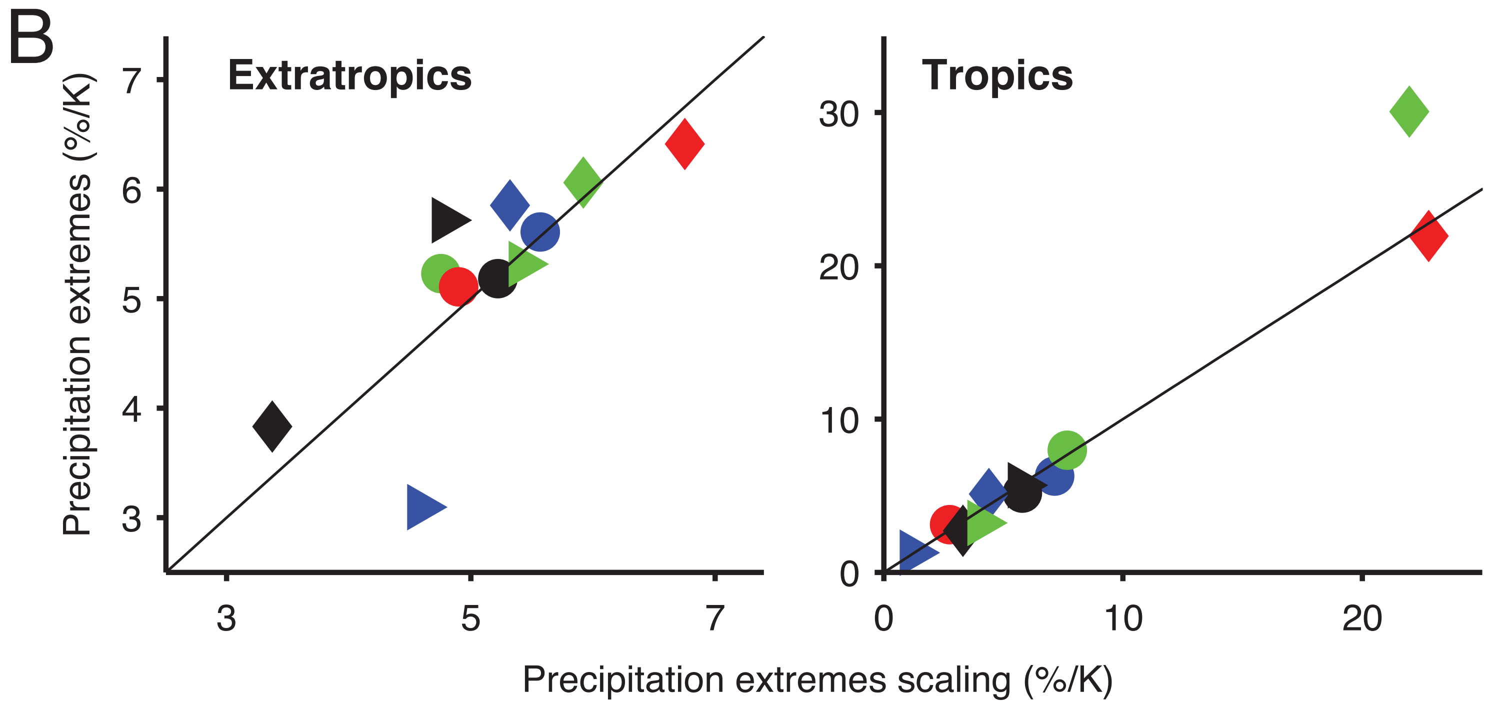
Scaling accurately reproduces simulated precipitation extremes in the current climate

Scaling requires knowledge of vertical velocity and temperature on days of precipitation extremes



Multi-model mean over 22 CMIP5 models. Scaling calculated using (5) (from Pfahl et al. 2015).

Scaling also explains changes in precipitation extremes across models



Changes in precipitation extremes and scaling (5) in CMIP3 models under the A1B scenario (from O'Gorman & Schneider, 2009)

Thermodynamic and dynamic contributions to precipitation extremes

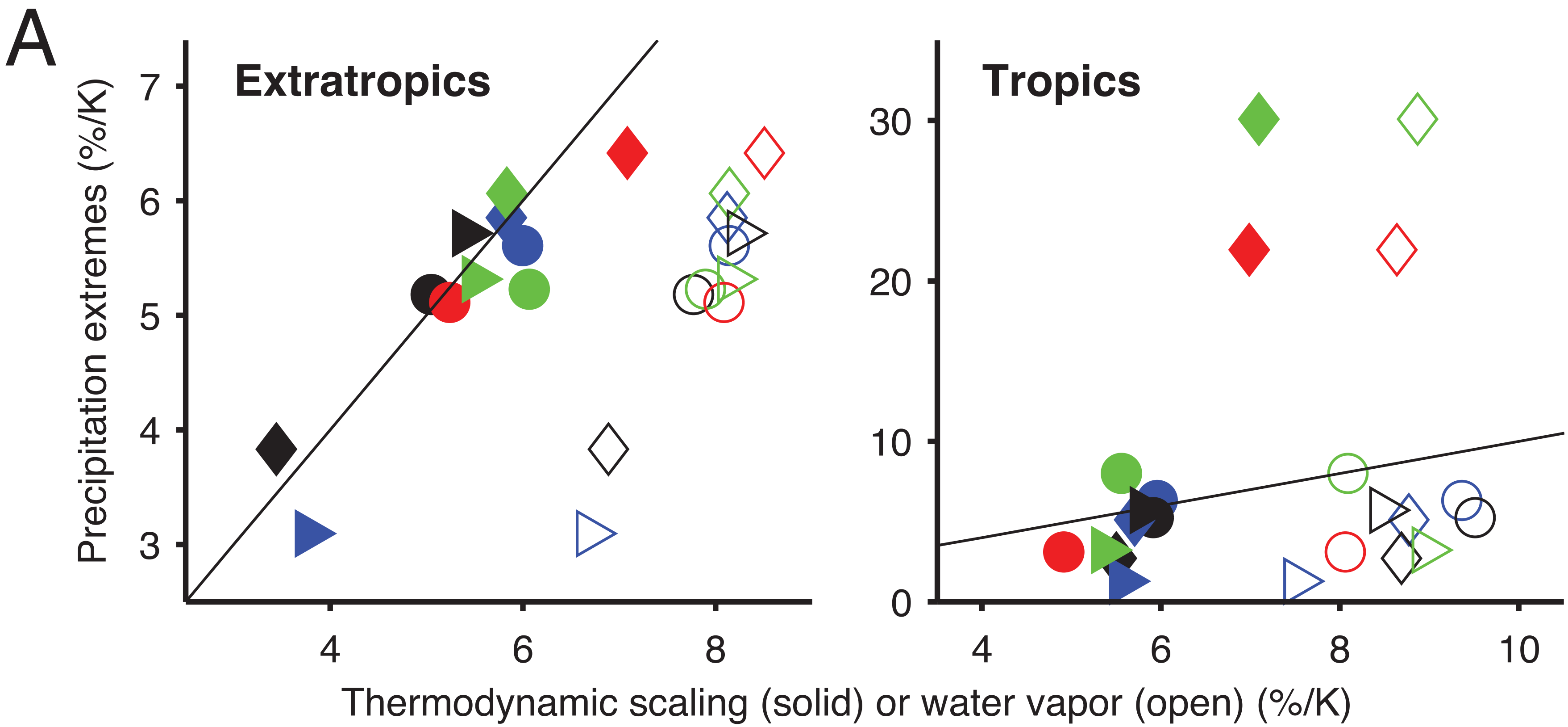
A thermodynamic scaling can be constructed by neglecting changes in vertical velocity

$$\delta P \approx \int \omega \frac{d\delta q_s}{dp}\bigg|_{\theta_e^*} \frac{dp}{g} + \int \delta\omega \frac{dq_s}{dp}\bigg|_{\theta_e^*} \frac{dp}{g}$$

thermodynamic

dynamic

Thermodynamic contribution dominates in the extratropics



Thermodynamic scaling calculated by neglecting ω in scaling (4).

Still requires knowledge of temperature during precipitation extremes (may be different to mean temperature).

(from O'Gorman & Schneider, 2009)

Summary so far

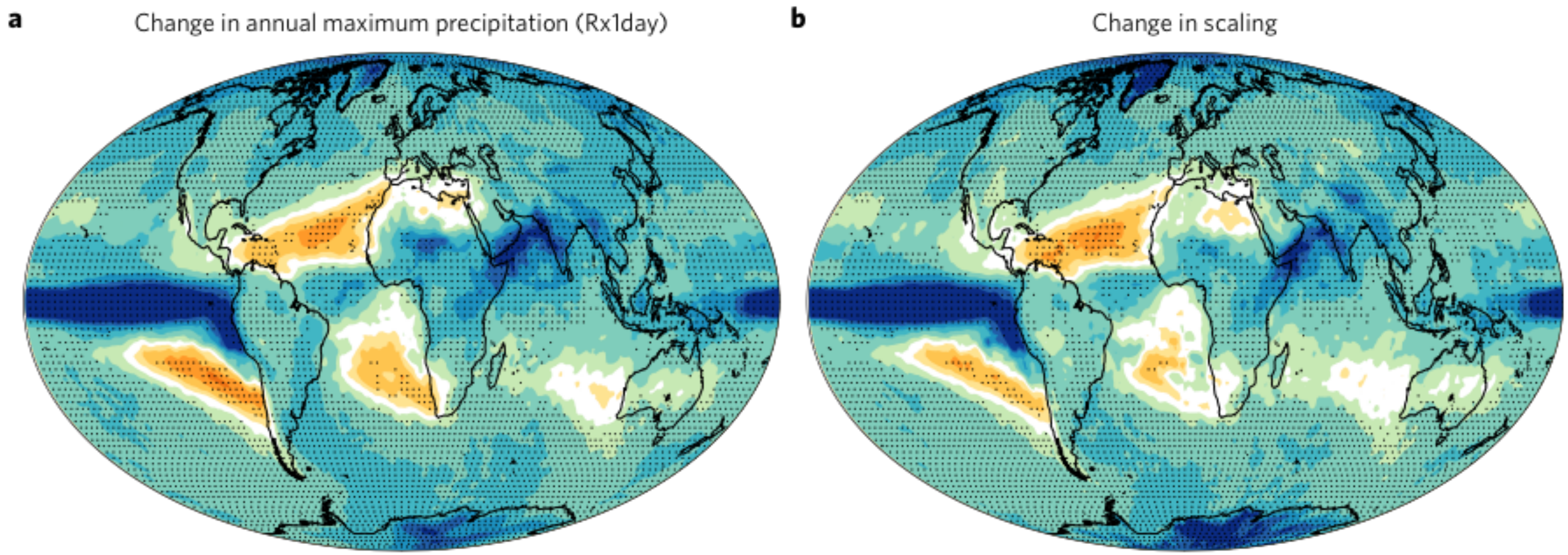
- Constructed scaling for daily precipitation extremes

- Indicates precipitation extremes increase at 4-8%/K if no dynamic or microphysical contributions

- GCM projections indicate microphysical contributions are small for daily extremes at the grid box scale, but dynamical effects may be large, particularly in tropical regions

- GCMs disagree strongly on dynamical contribution over the tropics as a whole

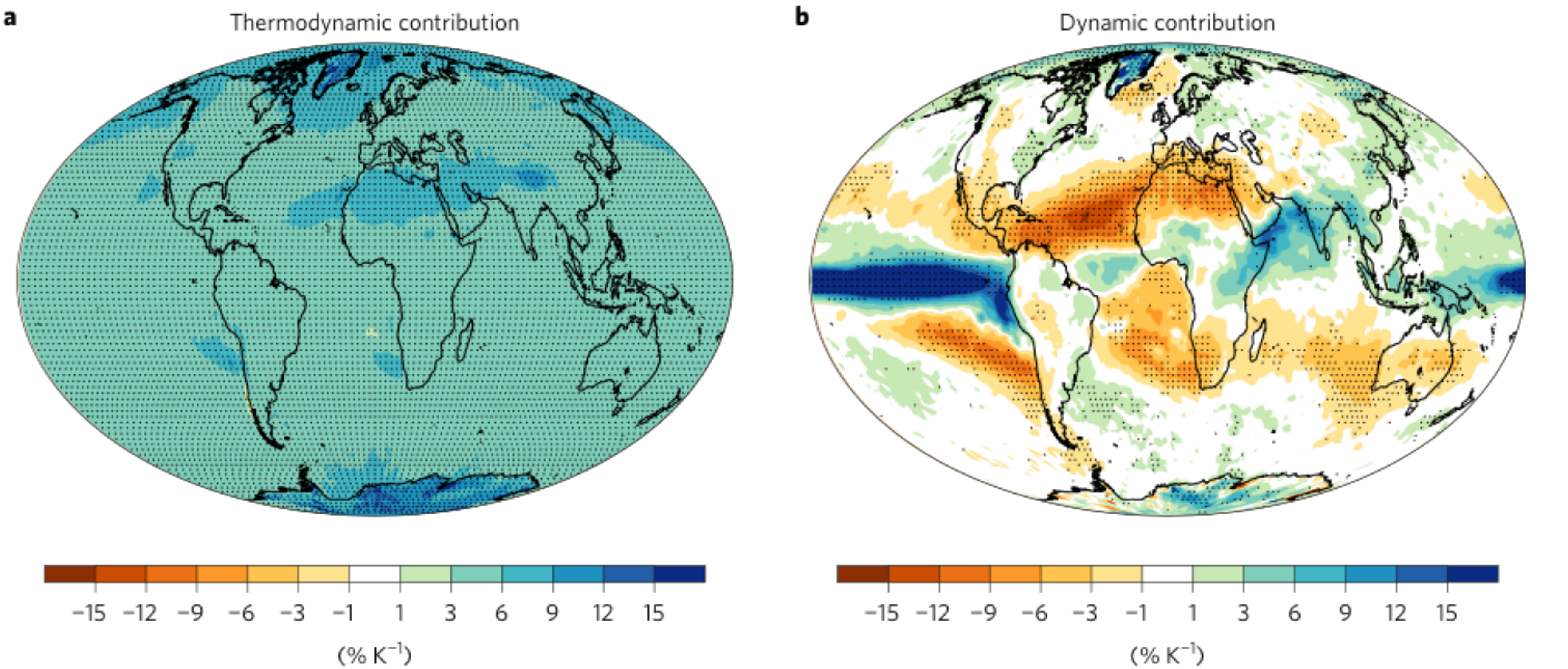
Spatial pattern of Precipitation extreme changes



Multi-model mean over 22 CMIP5 models. Scaling calculated using (5) (from Pfahl et al. 2015).

Spatial pattern determined by dynamic component

$$\delta P \approx \int \omega \frac{d\delta q_s}{dp}\bigg|_{\theta_e^*} \frac{dp}{g} + \int \delta\omega \frac{dq_s}{dp}\bigg|_{\theta_e^*} \frac{dp}{g}$$



Multi-model mean over 22 CMIP5 models. Scaling calculated using (5) (from Pfahl et al. 2015).

Robust decreases in precipitation regions over subtropical oceans

- Some agreement across models of negative dynamical contribution in subtropical regions

- Currently not well understood

- Starting point to examine quasi-geostrophic omega equation (e.g., Tandon et al., 2018)

$$\frac{N^2}{f^2}\nabla_h^2\omega + \partial_{pp}\omega = \mathcal{F}(\zeta, T, Q)$$

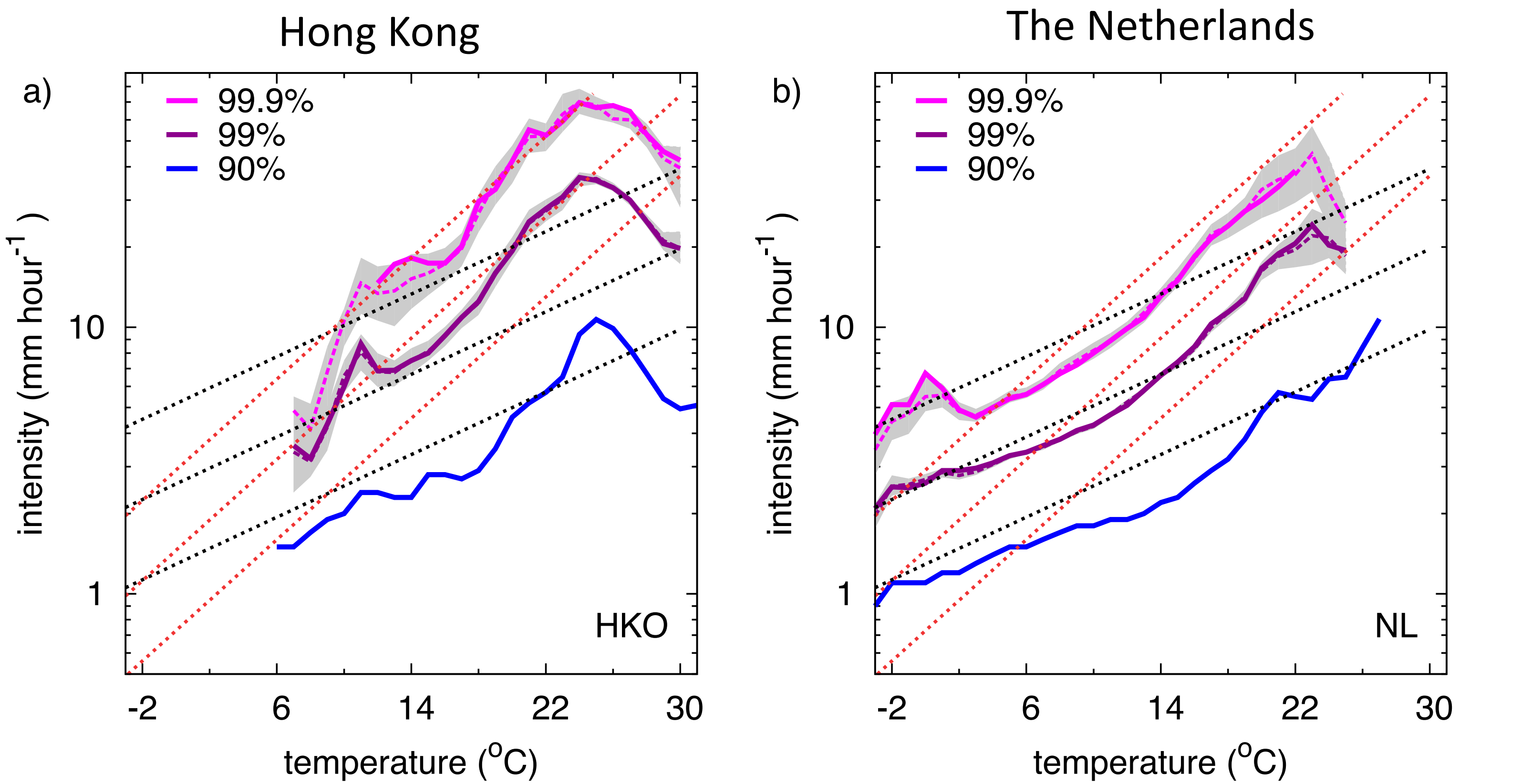
Convective-scale precipitation extremes

- Observations of precipitation extremes are typically based on gauges

- These are effectively point measurements and at a different spatial scale to GCM results

- In particular, gauges affected by small-scale convective precipitation extremes

Super Clausius-Clapeyron scaling in observed extremes



(from Lenderink et al., 2011).

Can these results be used to reason about global warming?

Temperature dependence of convective precipitation extremes

Simulating convective precipitation extremes requires cloud-resolving models

- Convective updraft velocity may increase with warming
  - Singh & O'Gorman (2015)
  - Loriaux et al. (2013)

- Efficiency may depend on warming
  - Singh & O'Gorman (2014)

Cloud-resolved modelling evidence mixed

- Some studies find super-CC scaling in convective precipitation extremes (Kendon et al., 2014) while others do not (Ban et al., 2015)
- May depend on region or model
- Clear that observed scaling rates may not be good indication of future changes (Bao et al., 2017)

Conclusions

- Precipitation extremes increase with warming in climate simulations and in observed trends

- Rate of increase varies across models and across regions because of uncertain dynamical contribution

- Better constraints on changes to precipitation extremes require understanding of the factors that determine the large-scale vertical velocity field

- Convective-scale precipitation extremes pose additional challenges

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Extra slides

Mean precipitation

In steady state, the global-mean atmospheric energy budget may be written,

$$F_{surf} - F_{TOA} = LHF + SHF$$

Radiative fluxes

Turbulent fluxes

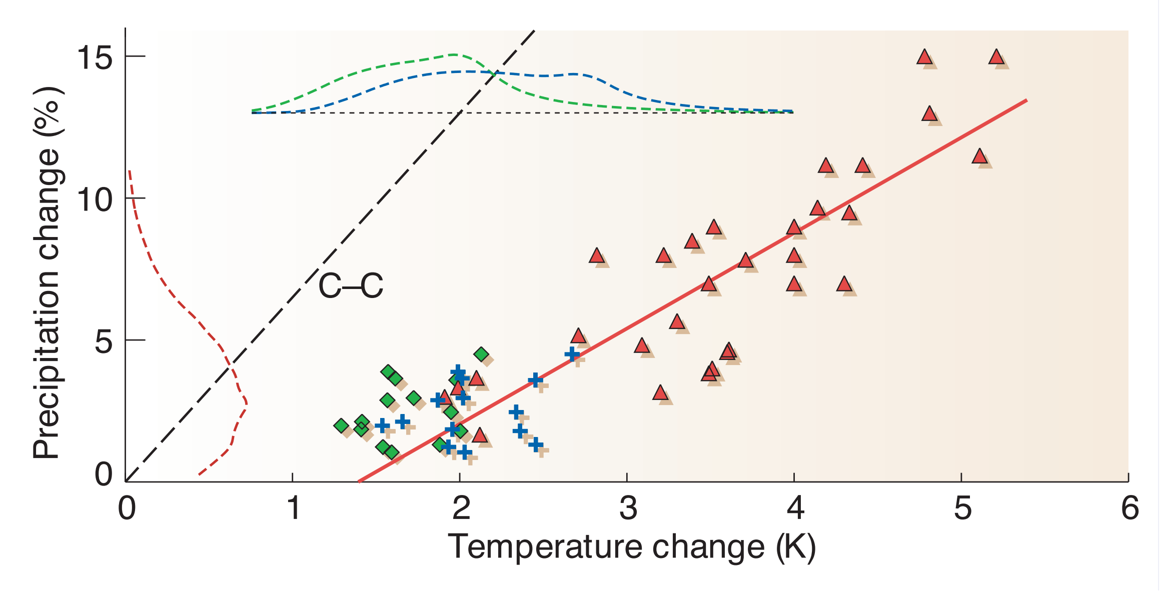
This states that the turbulent fluxes balance the radiative cooling rate Q.

On a global scale, latent heat fluxes dominate and we have

$$P \approx Q$$

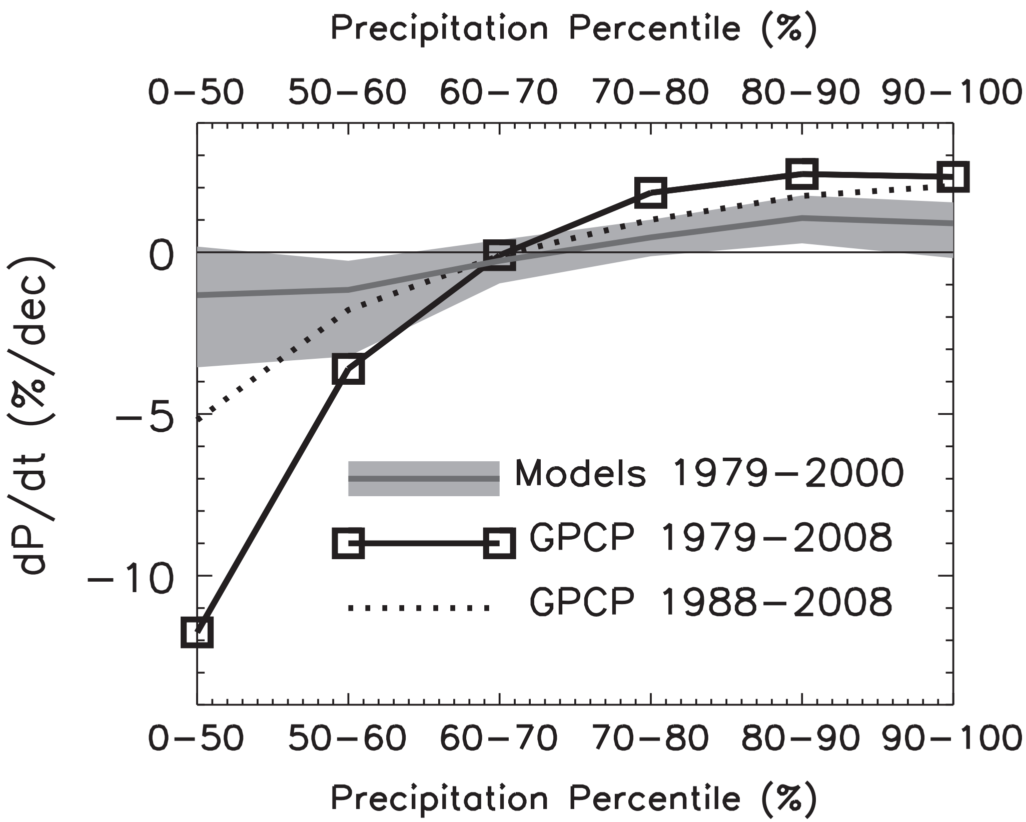
Mean precipitation

The radiative cooling rate increases under greenhouse gas forcing at a rate of roughly 2-3% per Kelvin



Increase in global-mean precipitation plotted against increase in global-mean temperature for doubling of CO_2 *(From Allen & Ingram, 2003)*

But this does not constrain precipitation extremes!



Time rate of change in tropical (30S-30N) precipitation rate as a function of percentile for observations (GPCP) and AMIP3 models (from Allan et al. 2010)