Workshop on Climate Studies, International Centre for Theoretical Sciences, Bengaluru, 1-3 March, 2022.

Aerosol Influences on Temperature, Cloud Properties and Rainfall over India

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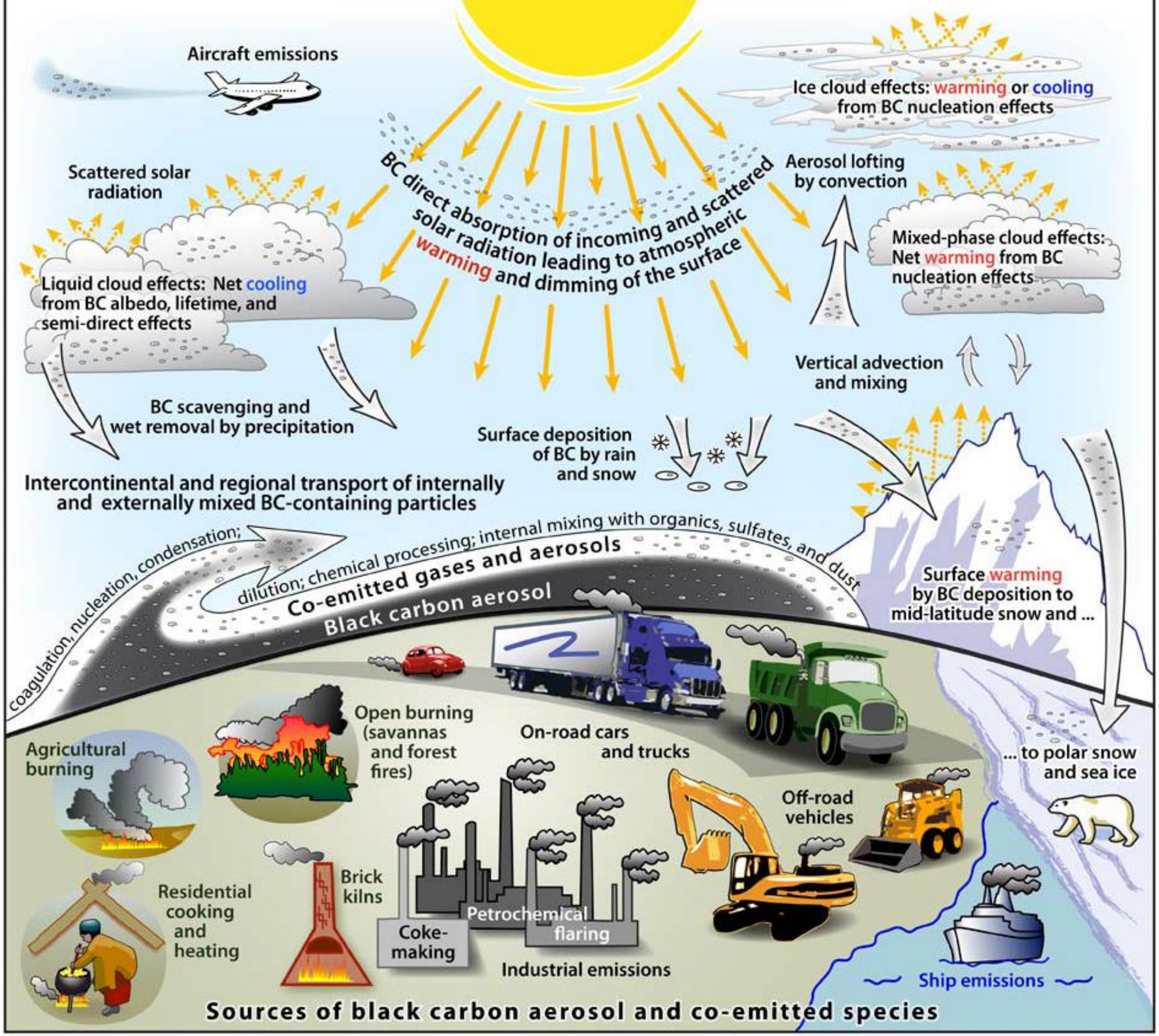
* Deceased on October 29, 2020, from COVID related complications in Kolkata.
** Deceased on February 9, 2022, following cardiac arrest in Jodhpur.

Aerosol Processes in the Climate System



C. Venkataraman, chapter lead author, “BC Global Emission Magnitudes and Source Categories”
In: Bond et al. 2013, JGR, “Bounding the climate effects of black carbon: A scientific assessment.”

Observational data and model simulation approaches

Observational studies to statistically extract aerosol effects on temperature, clouds and rainfall:

- Absorbing aerosol enhancement to extreme temperature events.
- Causal effects of aerosol influence on rainfall extremes
- Aerosol alteration of cloud microphysics / macrophysics.

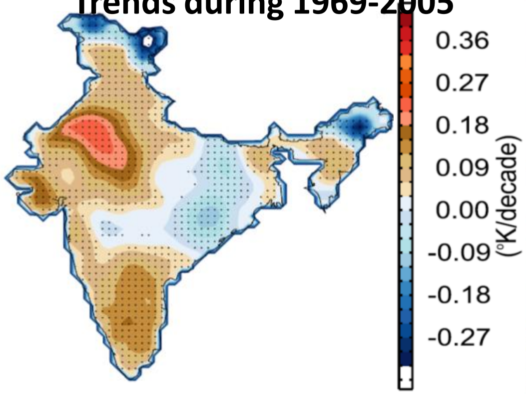
General circulation modelling to understand mechanisms of aerosol influences.

- Mechanisms of absorbing aerosol modulation of high temperature extremes.
- Disentangling aerosol & SST effects on mean rainfall.
- Aerosol perturbation of spatial heterogeneity in stratiform rainfall formation processes.
- Mechanisms underlying aerosol modulation of rainfall extremes.

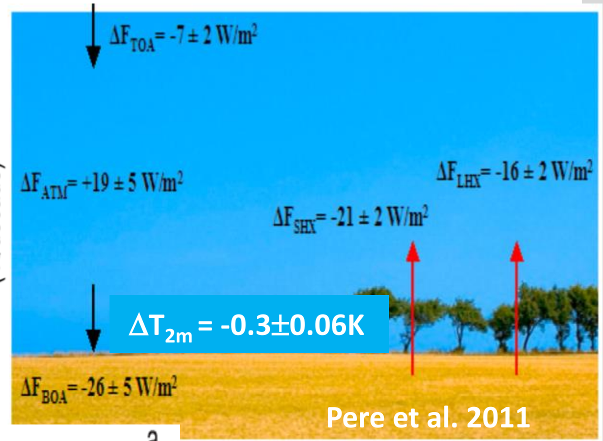
Temperature extremes

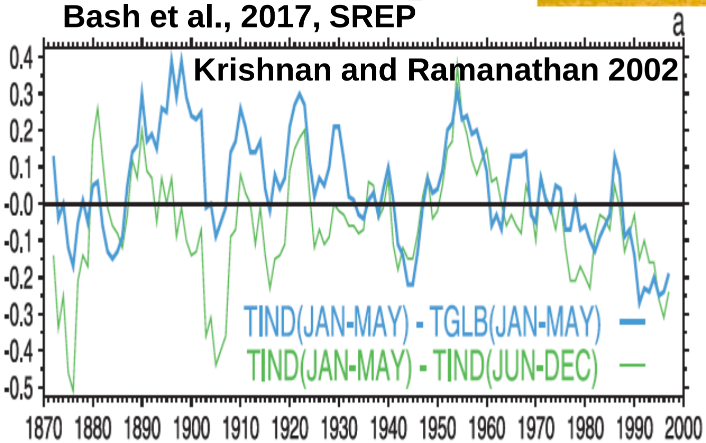
Hot extremes and aerosol-surface temperature links

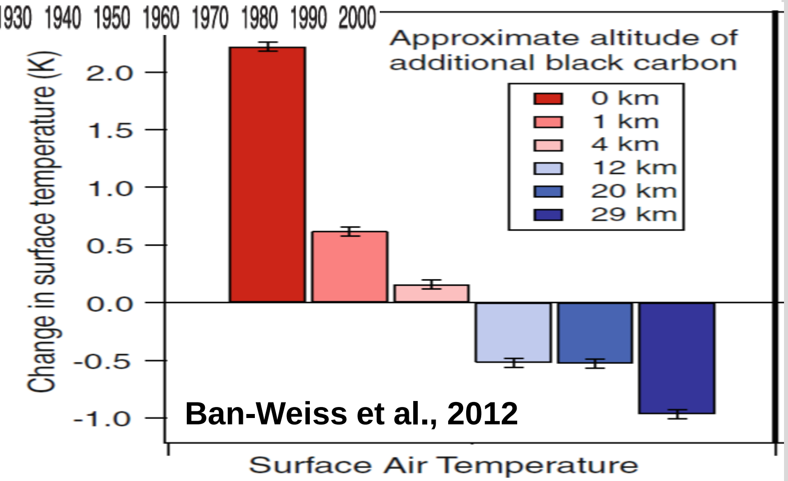
IMD Surface temperature Trends during 1969-2005



Bash et al., 2017, SREP

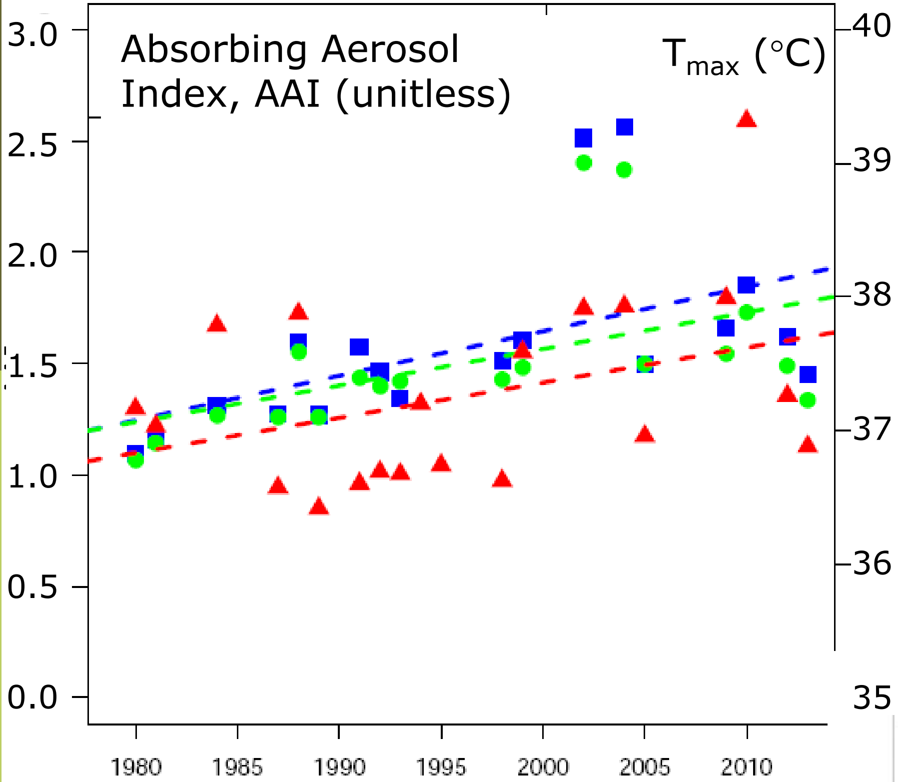






- **North and Central India heat waves:** linked with North-Atlantic blocking related anticyclonic conditions (Ratnam et al., 2016); subsidence over north India, depleted soil moisture, reduced precipitable water content, clear skies (Rohini et al., 2016).
- **Surface temperature cooling results from aerosols, largely sulphates,** in many world regions (Koch et al. 2009; Kloster et al. 2009; Zanis et al. 2012; Mickley et al. 2012; Pere et al. 2011)
- **Cooling trends in the dry season**, were linked to absorbing aerosols, with caveat that local cooling must be balanced by non-local warming. (Krishnan & Ramanathan, 2002).
- Ban-Weiss et al (2012) showed **low-altitude black carbon resulted in surface warming** through diabatic heating.

Absorbing aerosols intensify heat waves: Observational study

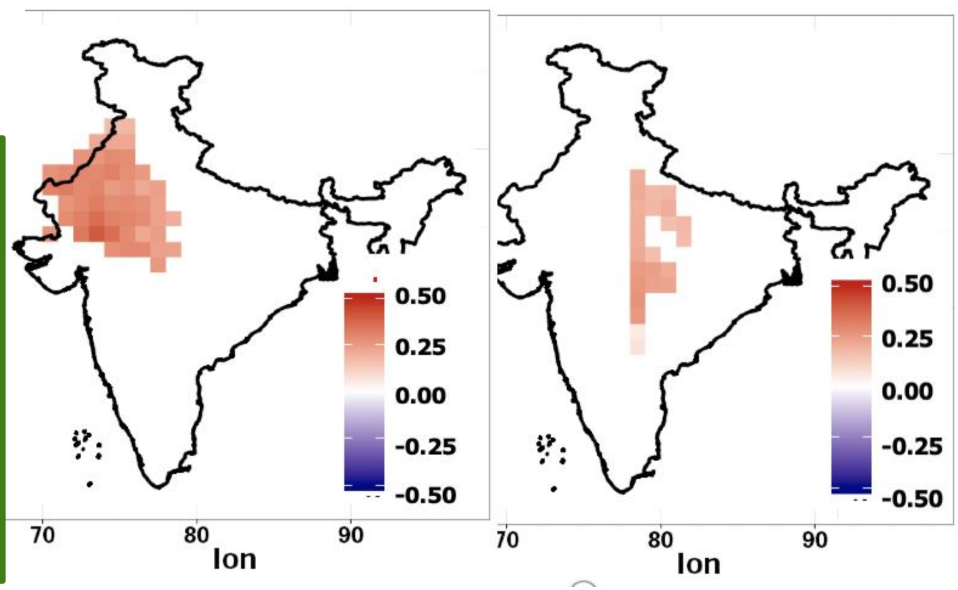


Between 1979-2013 (35 years), heat-waves were recorded in 20 years during MAMJ.

Increasing trends are found:

- In NW India of T_{max} (0.04 °C/y)
- Absorbing aerosol index (AAI, 0.03/y) over NWI and Central India.

AAI increases imply increases in black carbon / brown carbon and dust levels.

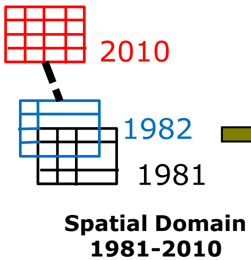


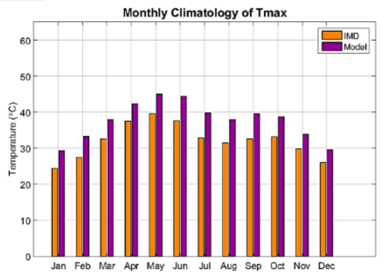
- Cross correlation analysis, T_{max} in NW India is influenced by AAI in both NW India and central India on the temporal scale of heat-wave events.
- For 9 years, local aerosol causality AAI-NW to Tmax-NW; for 11 years, non local aerosol causality, AAI-CI to Tmax-NW.

Dave, P., M.Bhushan and C. Venkataraman (2019) *Atmos. Environ.*, https://doi.org/10.1016/j.atmosenv.2019.117237

Absorbing aerosols influence high temperature extremes in India: ECHAM6-HAM2 general circulation model

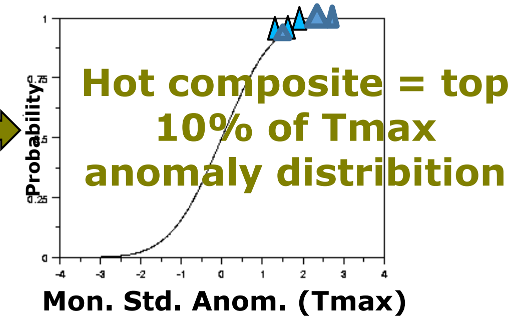
Methodology



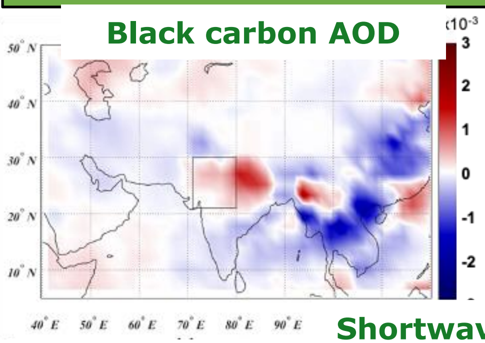


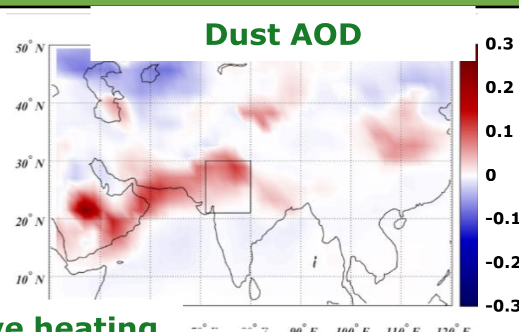
Monthly Std. Anomalies

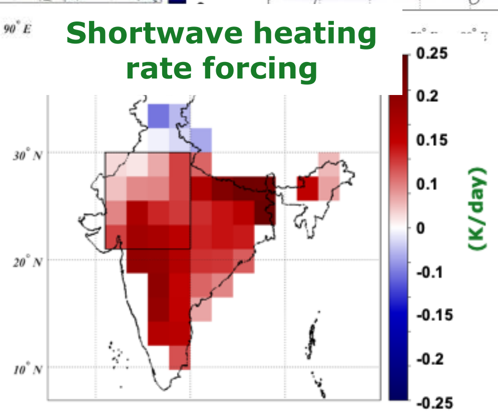
$$= \frac{T_{max\,yr}^{mon} - \overline{T_{max}^{mon}}}{Std.\,dev}$$



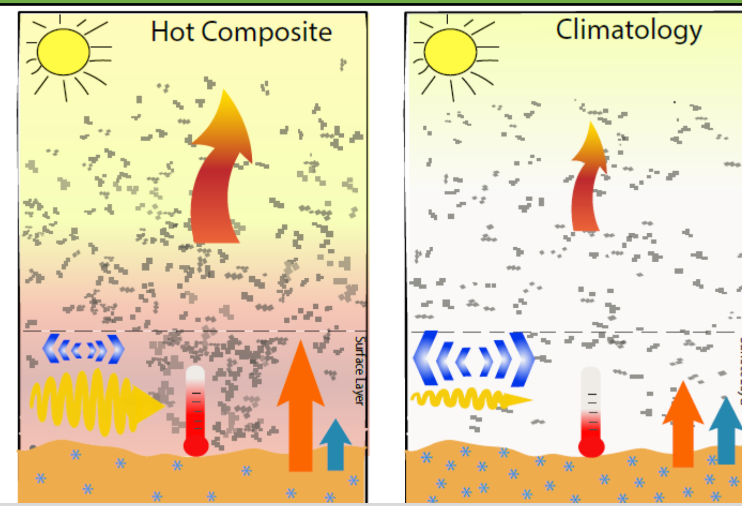
ΔHot composite – climatology (increased absorbing aerosols & heating rate)







Role of absorbing aerosols

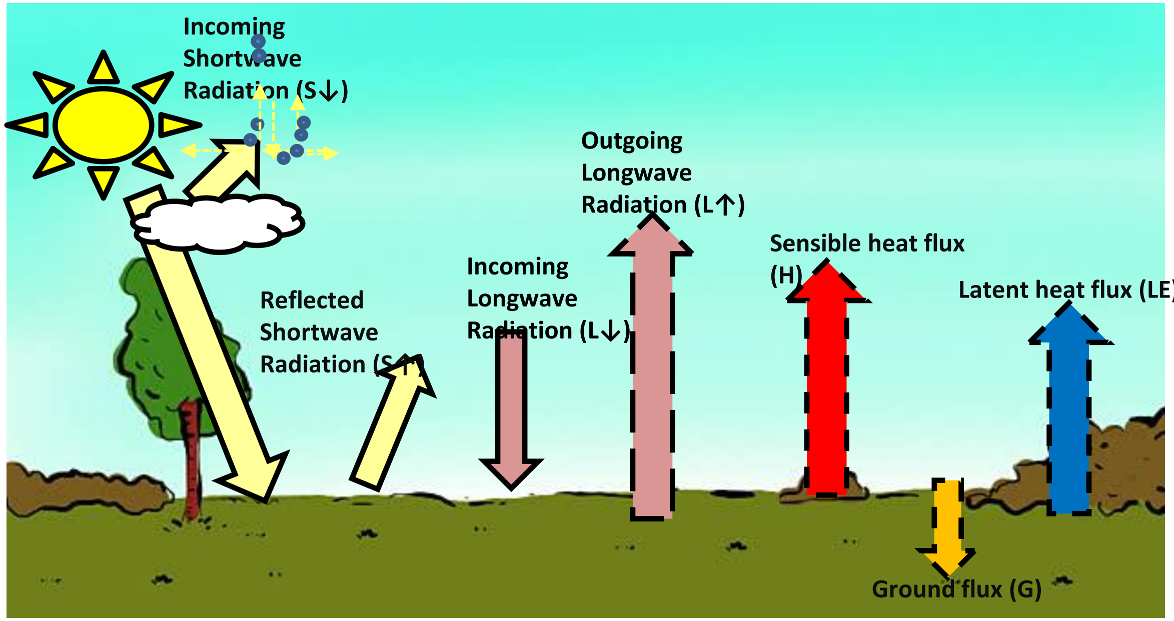


During hot extremes, higher surface temperatures consistent with:

- Increased black carbon, dust, radiative forcing (atm), SWHR forcing, sensible heat flux.
- Decreased single scatter albedo, latent heat flux.

Mondal, A, N. Sah, A.Sharma, C.Venkataraman & N.Patil (2020) Int. J. Climatol., https://doi.org/10.1002/joc.6783.

Land-atmosphere coupling: turbulent heat flux, vertical profiles in the surface layer



The surface energy balance can be written as:

$$C_L \frac{\partial T_s}{\partial T} = R_{net} + LE + H + G$$

$\boldsymbol{Ts}$, surface soil temperature, $\boldsymbol{C_L}$ heat capacity of the layer, ***H*** is the sensible heat flux, ***LE*** the latent heat flux, ***G*** is the ground heat flux and $\boldsymbol{R_{net}}$ the net radiation:

$$R_{net} = (1 - \alpha_s)R_{sd} + \epsilon R_{ld} - \epsilon\sigma {T_s}^4$$

$\boldsymbol{\alpha_s}$,$\boldsymbol{\epsilon}$ are surface albedo/emissivity, $\boldsymbol{R_{sd}}$,$\boldsymbol{R_{ld}}$ the downwelling shortwave/longwave radiation. The turbulent heat flux:

$$\overline{(w' {T_s}')}_s = -C_{T_s} |V_L| (T_L - T_s)$$

$\boldsymbol{C_{T_s}}$ is the heat transfer coefficient; T_L temperature at top of model surface atmospheric layer; $\boldsymbol{V_L}$ is horizontal wind vector. $\boldsymbol{C_{T_s}}$ is obtained from Monin-Obukhov similarity theory by integrating flux profile relationship over lowest model layer.

(Roeckner et al., 2003)

Land-atmosphere coupling: aerosol dry deposition velocity over different surfaces: e.g. snow surfaces

Observations of deposition flux (mass per unit area per time) are reconstructed from an ice core drill.

The BC deposition flux are overestimated by models:

- **Menegoz et al., 2014** - 16 times compared to observations (*model:* 53 mgm^{-2} yr^{-1}, *obs:* 3.2 mgm^{-2} yr^{-1})
- **Sharma et al., 2021** – 5 times compared to observations (*model:* 16 mgm^{-2} yr^{-1}, *obs:* 3.2 mgm^{-2} yr^{-1})

Dry deposition flux (F_c) = f(aerosol concentrations, C_z; and deposition velocity (V_d):

$$\boldsymbol{F_c = C_z V_d}$$

The deposition velocity $v_{d,k}$ of the kth moment is given by:

$$\boldsymbol{V_{d,k} = \frac{1}{r_a + r_{b,k} + r_a r_b V_{s,k}} + V_{s,k}}$$

Where r_a is the aerodynamic resistance, $r_{b,k}$ the quasi-laminar layer resistance and $V_{s,k}$ the sedimentation velocity ($\mathbf{V_{s,k}} = \boldsymbol{\frac{2}{9}\frac{r^2 \rho g Cc}{\mu}}$; C_c = slip correction coefficient; μ is viscosity of air)

Aerosol dry deposition velocity = f (friction velocity)

r_a is calculated from the roughness length z_0 and the boundary layer stability as below:

$$r_a = \frac{1}{u_* \, k}\left[ln\left(\frac{z}{Z_0}\right) - \emptyset\left(\frac{z}{L}\right)\right]$$

Where, u_* is the friction velocity, κ the von-Karmann constant of 0.4, z is the reference height (i.e., half of the lowest model layer height), Φ is a dimensionless stability term, and L the Monin-Obukhov-length.

$r_{b,k}$ is parameterized as following:

$$r_{b,k} = \frac{1}{\epsilon_0 u_* E_T}$$

Details of aerosol mechanics theory:

where E_T is the total collection efficiency and is parameterized as a series of collection efficiencies for each particle-surface interaction processes such as Brownian diffusion (E_B), interception (E_{IN})and impaction (E_{IM}) collection efficiency with the collecting surface i.e., $\mathbf{E_T = R\,(E_B + E_{IM} + E_{IN})}$

R = $e^{-St^{1/2}}$; $\quad$ $\mathbf{E_B = Sc^{-\gamma}}$; $\quad$ $\mathbf{E_{IM}} = \frac{St^2}{1+St^2}$; $\quad$ $\mathbf{E_{IN}} = \frac{1}{2}\left(\frac{D_p}{A}\right)^2$

where, R= correction factor; St = stokes number; Sc = schmidt number; A = characteristic radius of collectors, depends on land use category

Land-atmosphere coupling: Dust emission, threshold friction velocity underestimated

Friction velocity: (Assuming wind profile in IBL follows logarithmic law)

$$u_* = \frac{U(z) * k}{\ln(\frac{z}{z_{0s}})}$$

- **Friction Reynolds number: (empirical)**

$$B = 1331 D_p^{1.56} + 0.38$$

- **Threshold friction velocity:**

$$u_{*th} = g(B, D_p)$$

- **Threshold friction velocity (corrected for rough surface):**

$$u_{th*} = \frac{u_{th*}}{f_{drag}} \; [f_{drag} = f(z_0, z_{os})]$$

- **Threshold friction velocity (corrected for wet soil):**

$$u_{th*} = u_{th*}\sqrt{(1 + 1.21(w - w')^{0.68})} \; [\mathrm{w} > \mathrm{w}']$$

- **Threshold friction velocity (corrected based on region):**

$$u_{th*} = u_{th*} * \boldsymbol{nduscale}$$

✓ $U(z)$ is obtained from horizontal wind speed.

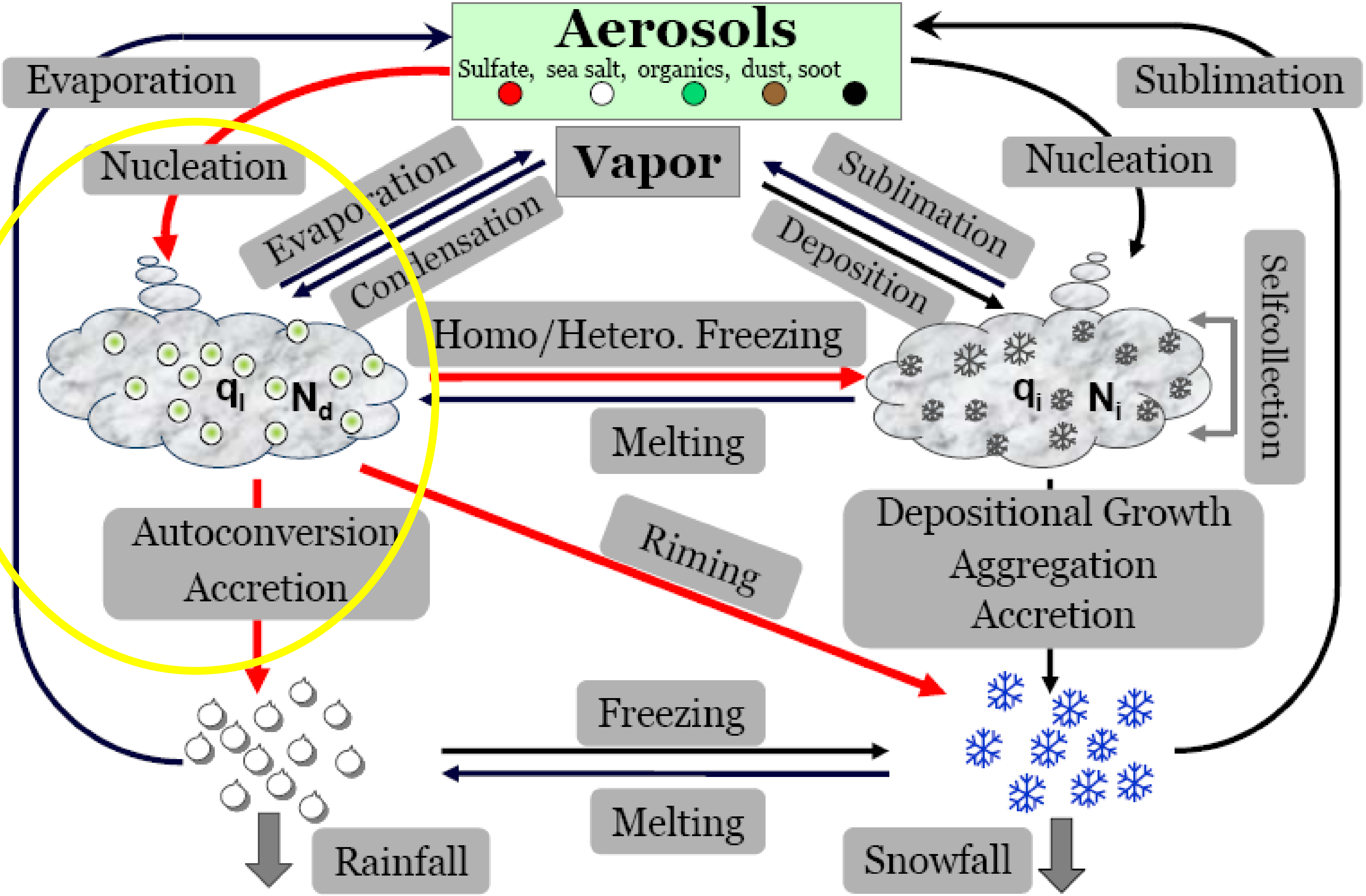
✓ If at a pixel, $u_* > u_{*th}$; it is considered as emitting pixel.

✓ Higher *nduscale* values hinders the emission.

Equations for calculation of emission *(Marticorena and Bergametti (1995) & Tegen et al. (2002)*

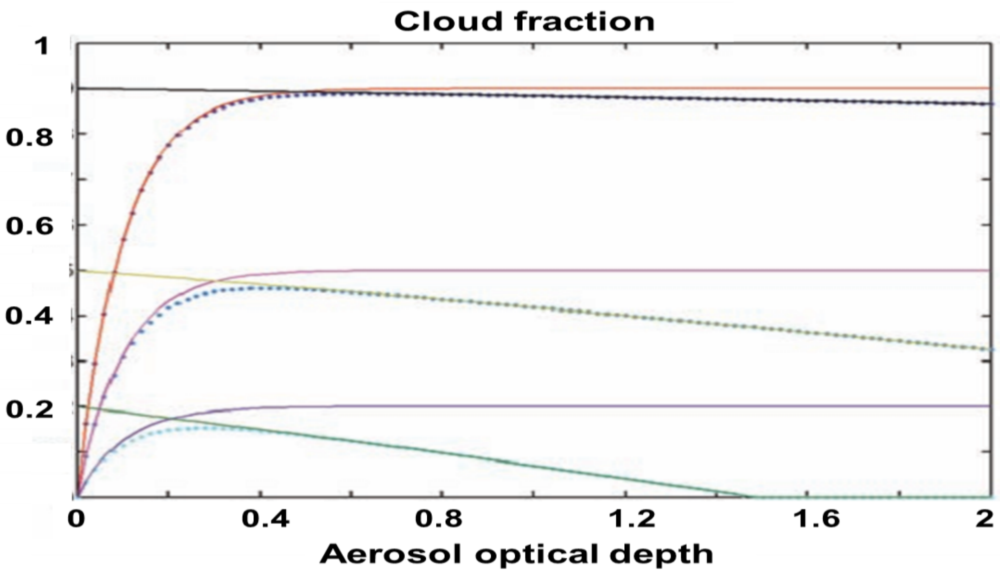
Cloud Microphysics and Macrophysics

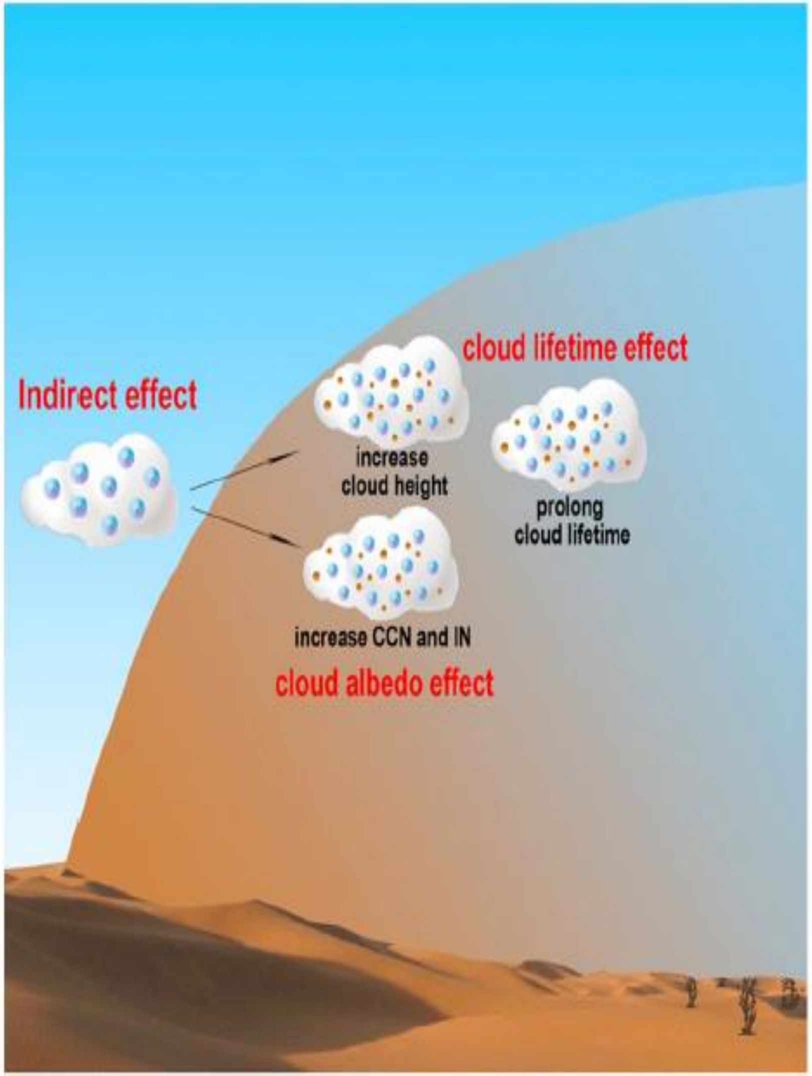
Cloud and rainfall processes



Source: Lohmann 2007

Cloud responses to aerosol radiative effects





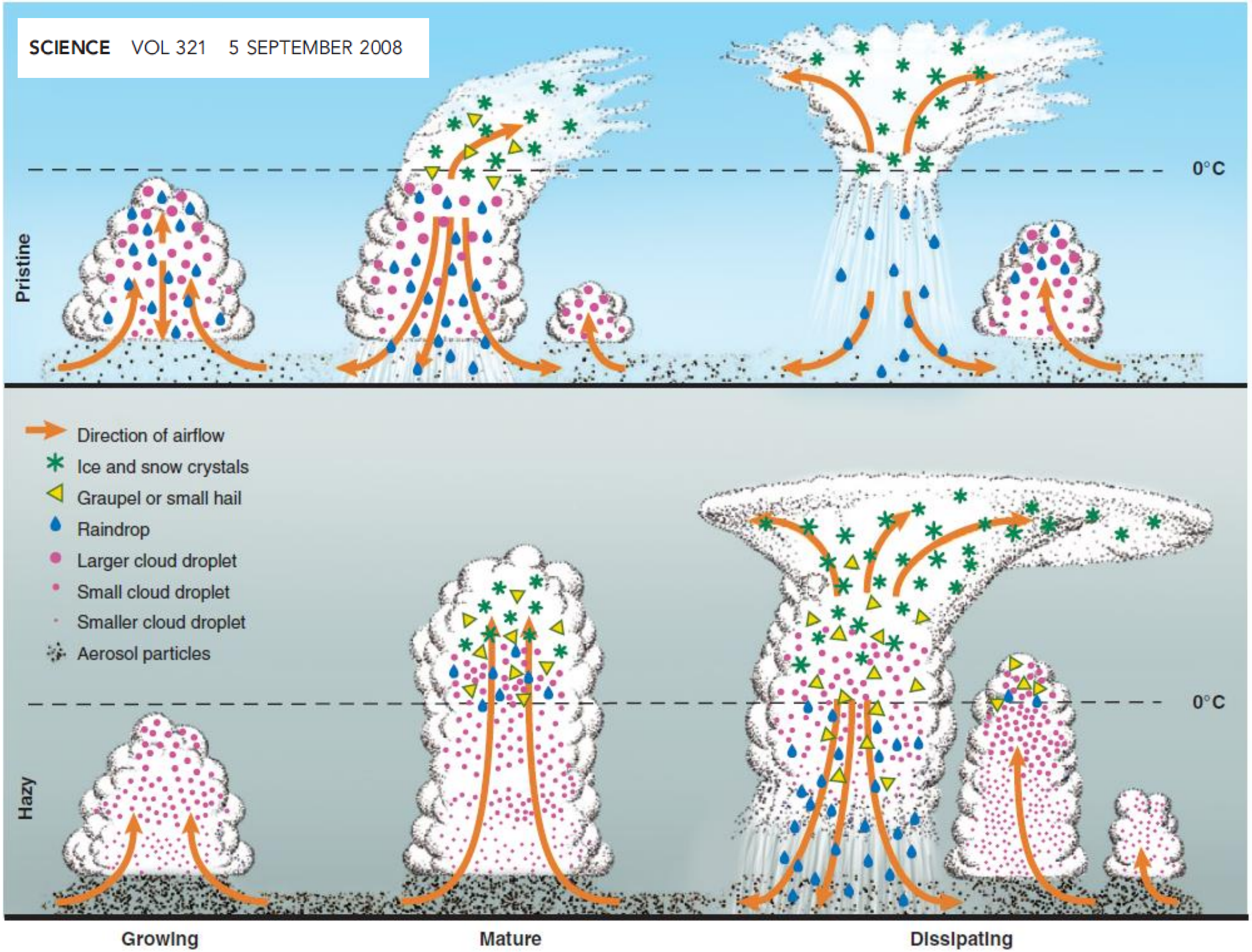
Absorbing aerosols stabilize near-surface atmosphere. Negative cloud feedback [*Koren 2008; Jacobson, 2002*].

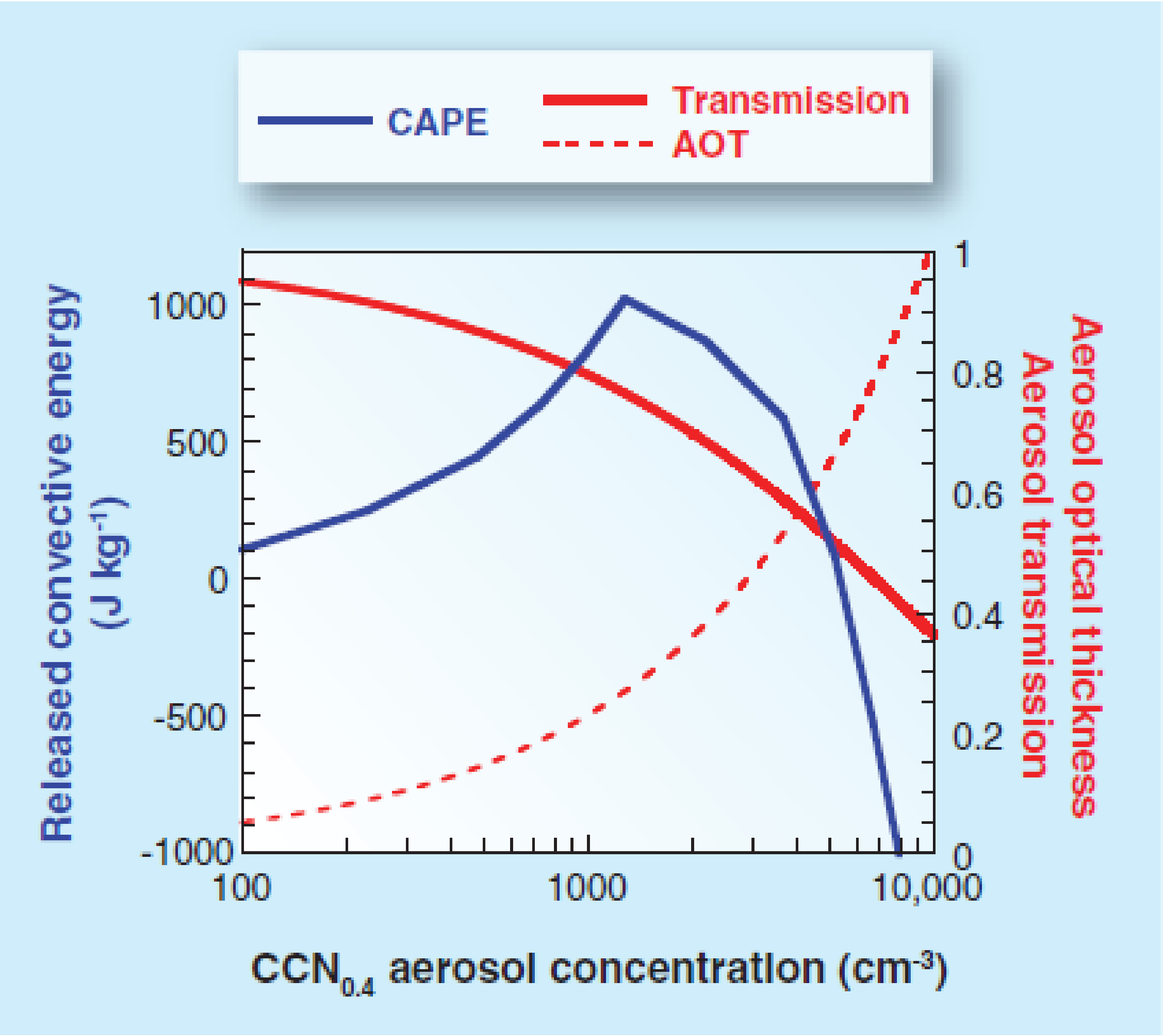
Inhibition of warm cloud development. Amazon biomass burning [*Ten Hoeve et al., 2011*].

Clouds contamination by dust, semi-direct effect, lower LWP and IWP. Taklimakan and East Asia [*Huang et al., 2006*; *2014*].

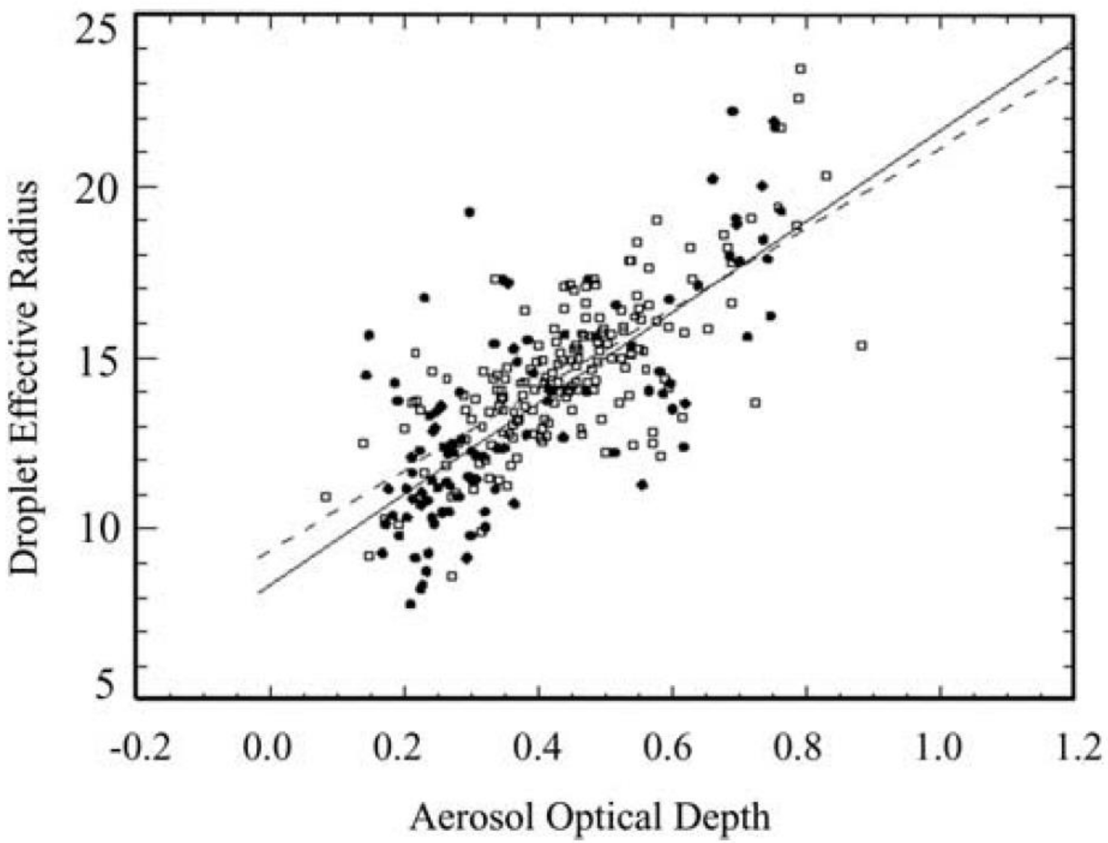
Flood or Drought: How Do Aerosols Affect Precipitation?

Daniel Rosenfeld,[1*] Ulrike Lohmann,[2] Graciela B. Raga,[3] Colin D. O'Dowd,[4] Markku Kulmala,[5] Sandro Fuzzi,[6] Anni Reissell,[5] Meinrat O. Andreae[7]




CAPE
Transmission
AOT
Released convective energy (J kg-1)
1000
500
0
-500
-1000
1
0.8
0.6
0.4
0.2
0
Aerosol optical thickness
Aerosol transmission
100
1000
10,000
CCN0.4 aerosol concentration (cm-3)

Cloud thermodynamic changes induced by aerosols

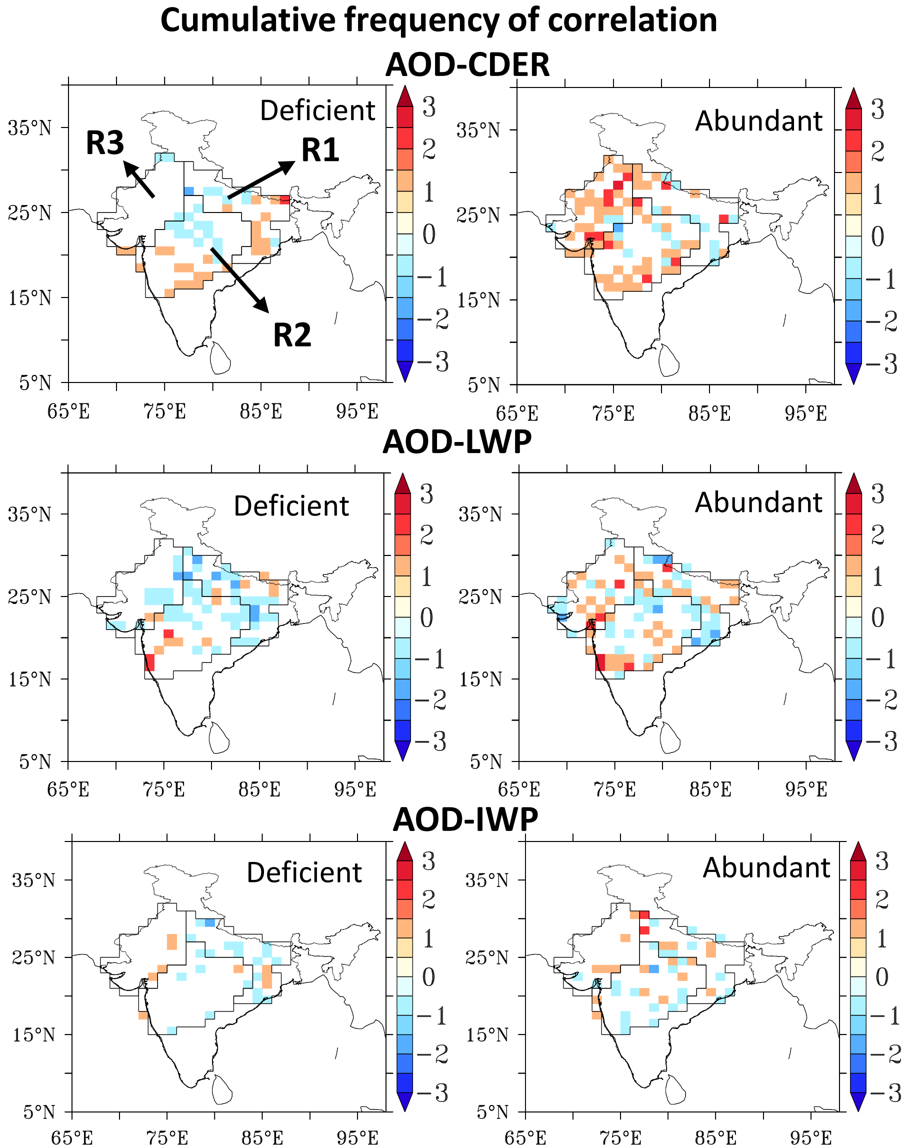


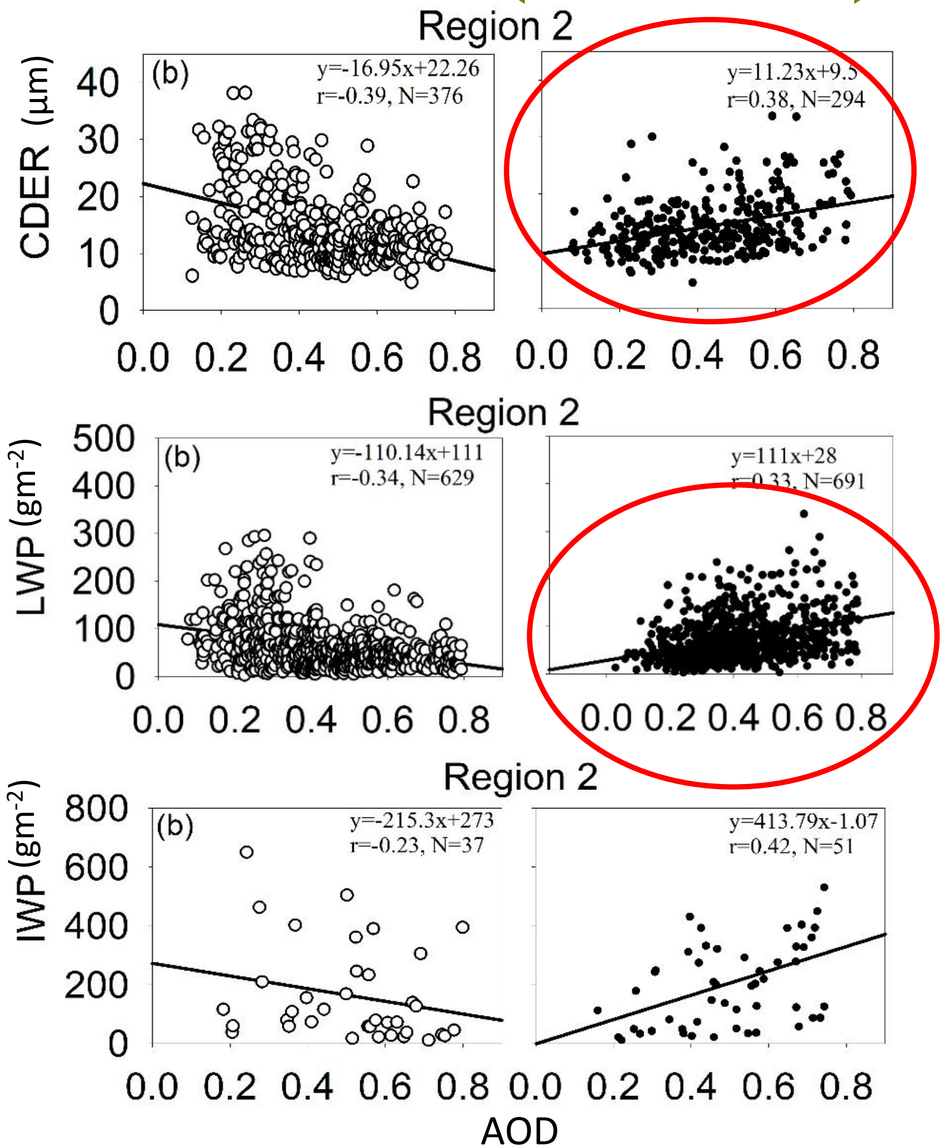
Positive AOD-CDER relationships

- Seen over warm, humid and polluted regions (Gulf of Mexico and South China Sea).
- Behaviour found in 2-D cumulus ensemble model, upon the introduction of slightly soluble organics (decrease in activated nuclei) or few giant CCN.

[*Yuan et al. 2008*]

Aerosol influence on CDER, LWP and IWP (2000-2009)





- Increased AOD correlate with decreases in cloud drop sizes (CDER), cloud liquid water path (CLWP) and ice processes (IWP) in deficient, but increases in abundant monsoon years.

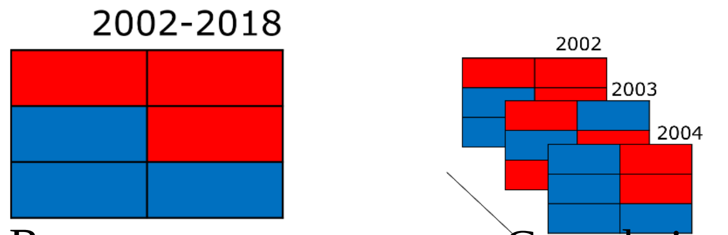
Patil, Dave, Venkataraman, 2017, Scientific Reports

Study ongoing (under revision): Evaluating the aerosol cloud interaction parameter *(aci)* over India (2001-2018)

Methodology

MODIS L3 AOD, CDER, TPW (2002-2018)

$$aci(r_e) = -\frac{\partial \ln CDER}{\partial \ln AOD}$$



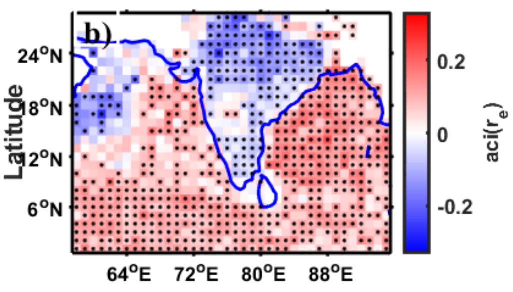
Red: +ve *aci*, (Twomey effect)

Cumulative frequency

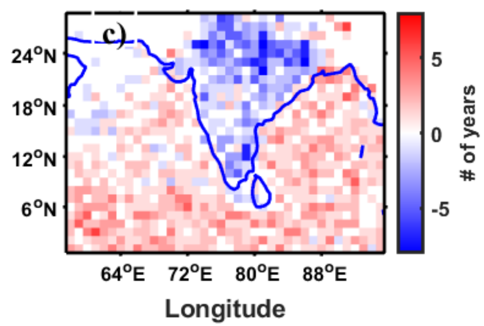
Mean $CDER_{highTPW}$ and $CDER_{lowTPW}$ for different AOD bins over land and ocean

Land-ocean contrast in *aci*

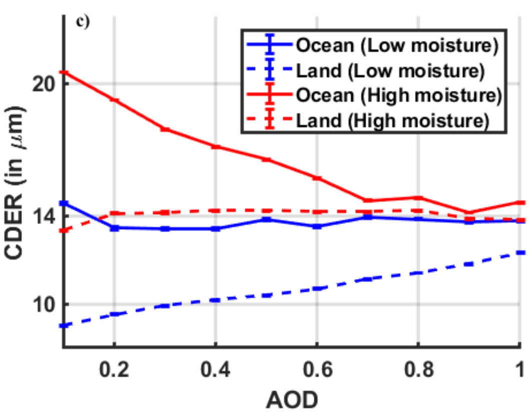
Spatial distribution of *aci*



Cumulative frequency of positive *aci*



Effect of moisture

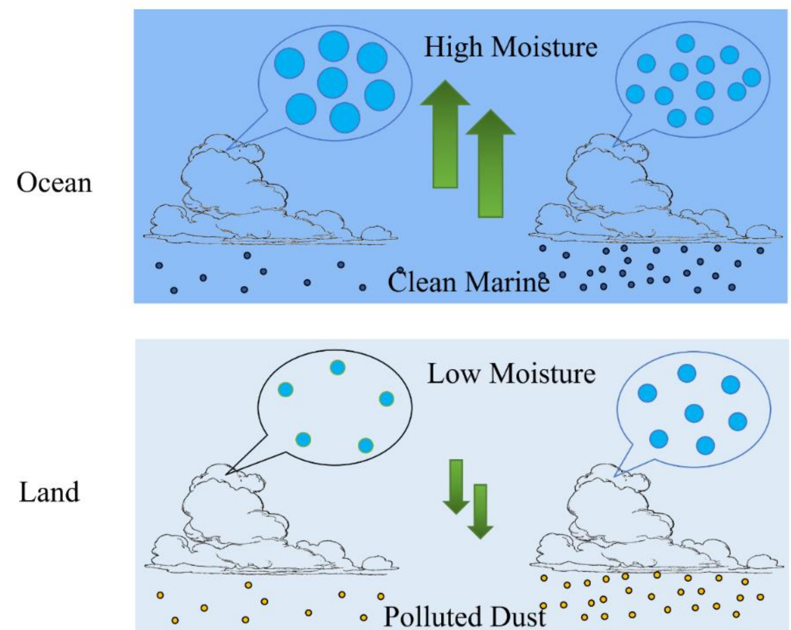


Negative *aci* (anti-Twomey effect) is consistent with

- polluted dust
- low moisture
- downdraft

over land.

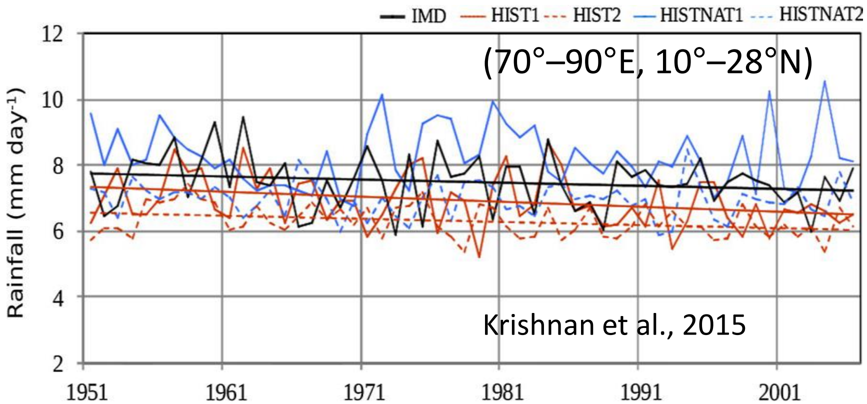
Effect of meteorology & aerosol type

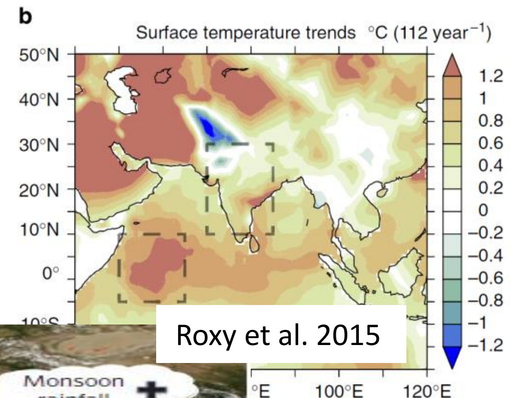


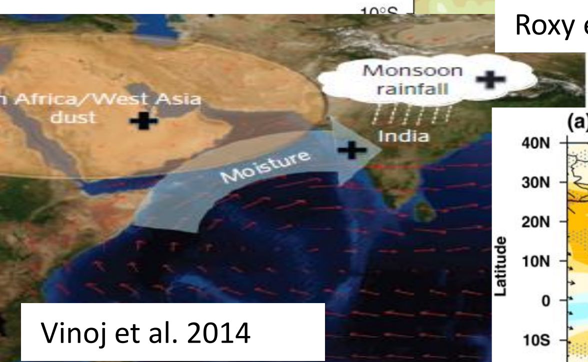
K. Muduchuru, M. Das et al. 2022, (under revision)

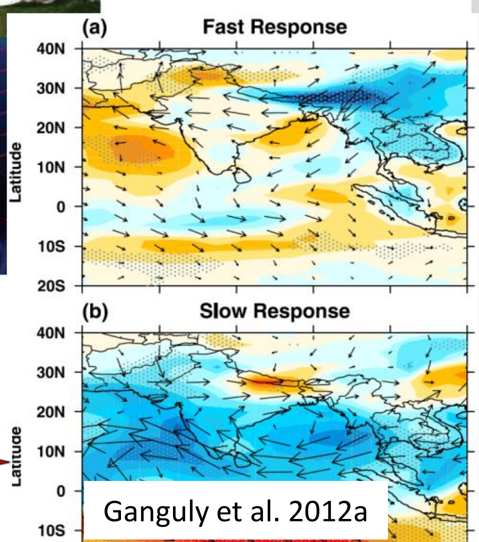
Mean rainfall

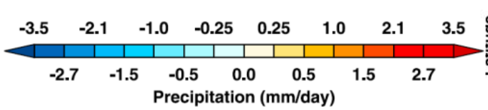
Influence of different forcing elements on the South Asian monsoon











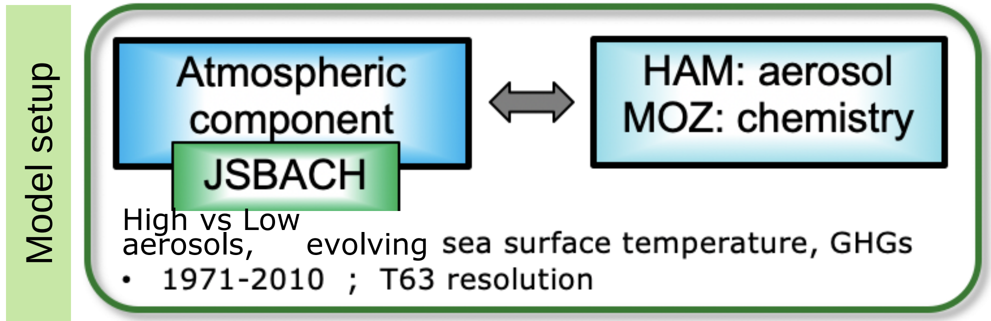
Monsoon rainfall weakening linked to an asymmetric inter-hemispheric energy balance change attributed to:

- **Aerosol enhancement** [e.g. *Bollasina et al. 2011; Ganguly et al. 2012a; Krishnan et al. 2015; Salzmann et al. 2014; Guo et al. 2016*].
- western **Indian ocean SST warming** [*Zhou et al. 2008; Roxy et al. 2014 and 2015*].

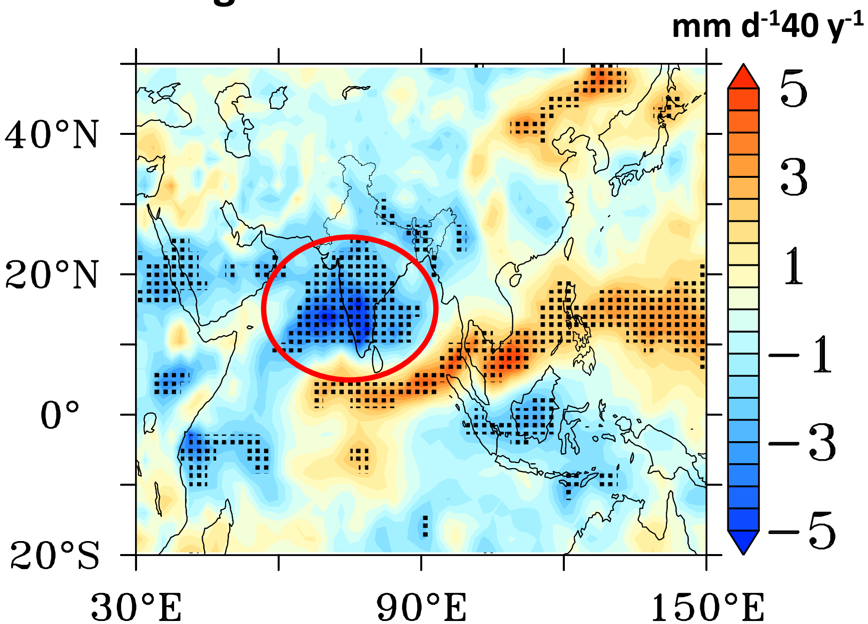
Short-term rainfall enhancement from fast responses to radiative effects of **non-local absorbing aerosols** *[Lau et al.2006, Vinoj et al., 2014].*

Thermodynamic, convection and circulation adjustments *influence rainfall weakening [Ramanathan et al.2005; Randles and Ramaswamy, 2008; Ganguly et al. 2012b].*

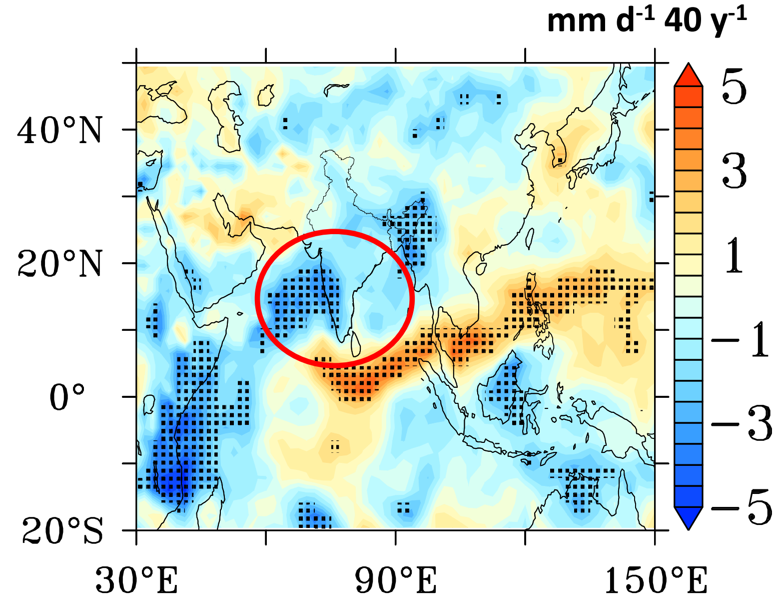
Aerosol influences on rainfall trends: ECHAM6-HAM2 GCM



High Aerosols – 2010 levels



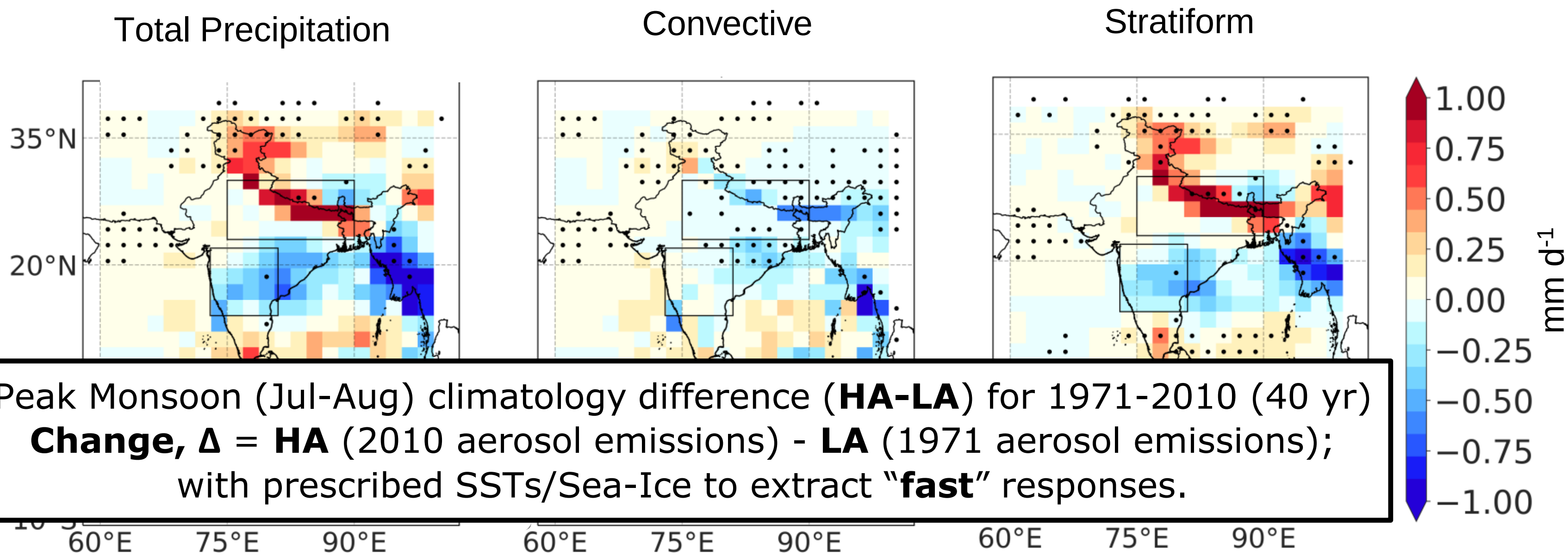
Low Aerosols – 1971 levels



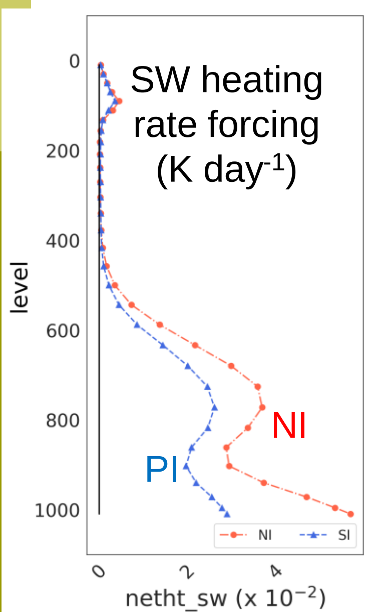
- Increased aerosol levels induce intensification of drying trend over peninsular India.
- Linked to changes in Hadley circulation, mesoscale convective activity, lower tropospheric stabilization

Patil et al. 2018, Climate Dynamics

Aerosol induced rapid adjustments on stratiform rainfall: ECHAM6-HAM2



Peak Monsoon (Jul-Aug) climatology difference (**HA-LA**) for 1971-2010 (40 yr)
Change, Δ = **HA** (2010 aerosol emissions) - **LA** (1971 aerosol emissions);
with prescribed SSTs/Sea-Ice to extract "**fast**" responses.

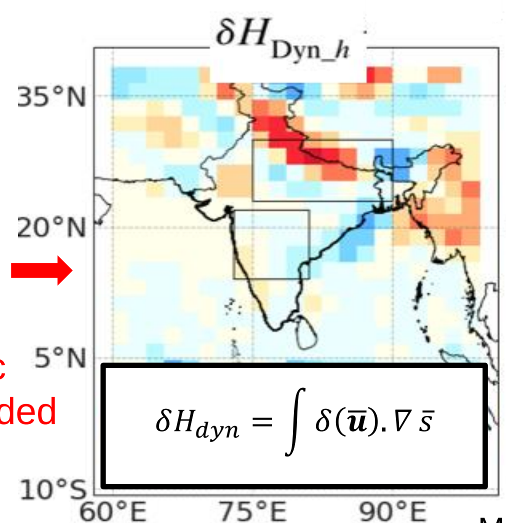


Energetic Constraints

Global: Rainfall change is balanced by diabatic cooling

$$\boxed{L_c \delta P = \delta Q - SH} + \boldsymbol{\delta H} \rightarrow$$

Regional: Additional dry static energy divergence term is added to balance rainfall change



$$DSE: \; s = gz + C_p T$$

Absorbing aerosols induce fast adjustments
→ Larger increases of SW heating rate forcing over NI
→ Stabilization over NI
→ Induce fast adjustments

Changes in regional **horizontal wind** changes ($\boldsymbol{\delta H_{dyn}}$) → moisture increases (NI); decreases (PI) → St. Rainfall changes

Muduchuru & Venkataraman, 2021, Climate Dynamics

Stratiform and convective parameterization schemes

$$\frac{dq_v}{dt} = R(q_v) - \underbrace{b\left(Q_{cnd}^c + Q_{dep}^c\right)}_{\text{Sink}} + (1-b)\left(\underbrace{Q_{evp}^c + Q_{sub}^o}_{\text{Sink}} - \underbrace{Q_{cnd}^o - Q_{dep}^o}_{\text{Source}}\right),$$

Cloud water increase → rainfall process cloud cover scheme

$$\frac{dq_l}{dt} = R(q_l) + b\left(\underset{\star}{Q_{cnd}^c} - \underbrace{\underset{\star}{Q_{aut}^c} - Q_{racl}^c - Q_{sacl}^c - Q_{frc}^c - Q_{frh}^c - Q_{frs}^c}_{\text{Sink}} + \underbrace{Q_{mlt}^c\right) + (1-b)Q_{cnd}^o}_{\text{Source}},$$

$$\frac{dN_l}{dt} = R(N_l) + \underset{\star}{Q_{nucl}} - \underbrace{\underset{\star}{Q_{autn}} - Q_{mlt} - Q_{agg} - Q_{saci}}_{\text{Sink}}$$

$\star \rightarrow$ Represents interaction with CCN (activated aerosols)

Cloud Fraction for Stratiform clouds →

$$b = \int_{r_s}^{\infty} G(r_t)\,dr_t$$

$$r_t = r_v + r_l + r_i$$

$v - water\ vapor$
$l - cloud\ liquid$
$i - cloud\ ice$

ECHAM6-HAM2: two-moment scheme for cloud microphysics: prognostic q_v, q_l, N_l, q_i, N_i

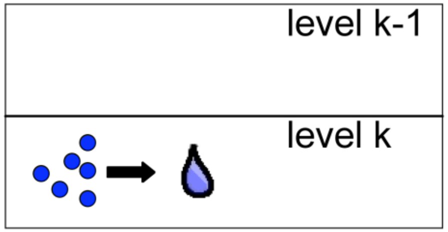
In ECHAM, the convective parameterization provides only for vertical transport and detrainment of cloud water.
→ Convective clouds have an “indirect” effect on radiation only through detrainment of liquid water to the stratiform scheme.

$$\left.\frac{\partial q_l}{\partial t}\right|_{\text{auto}} = -\gamma_l 1350\ q_l^{2.47}\ N_l^{-1.79}$$

Cloud water

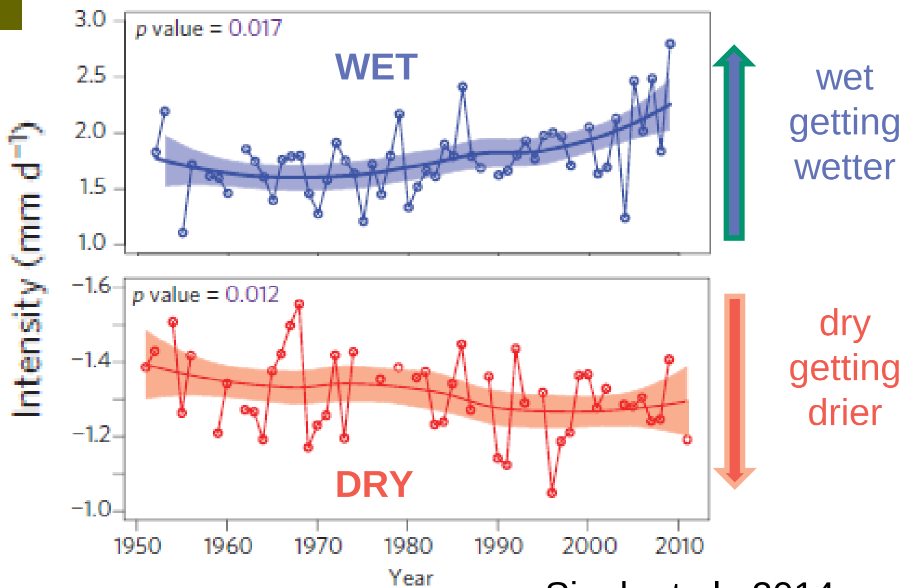
Cloud number conc.

Khairoutdinov and Kogan, 2000

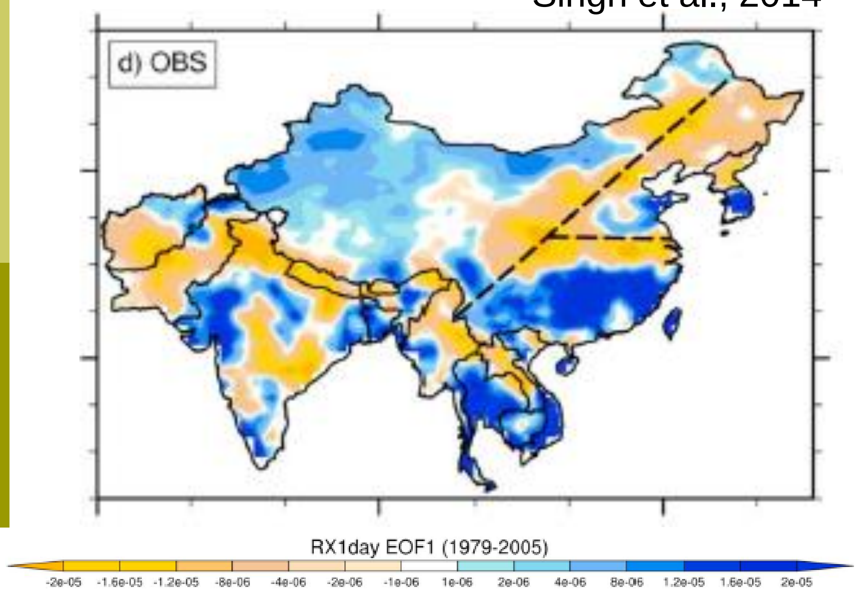


Extreme rainfall

Extreme rainfall over the Indian "core" monsoon region



Singh et al., 2014



Lin et al., 2018

Strong dependence of extreme daily characteristics to the forcing agent *(Lin et al., 2016; Sillmann et al., 2019).*

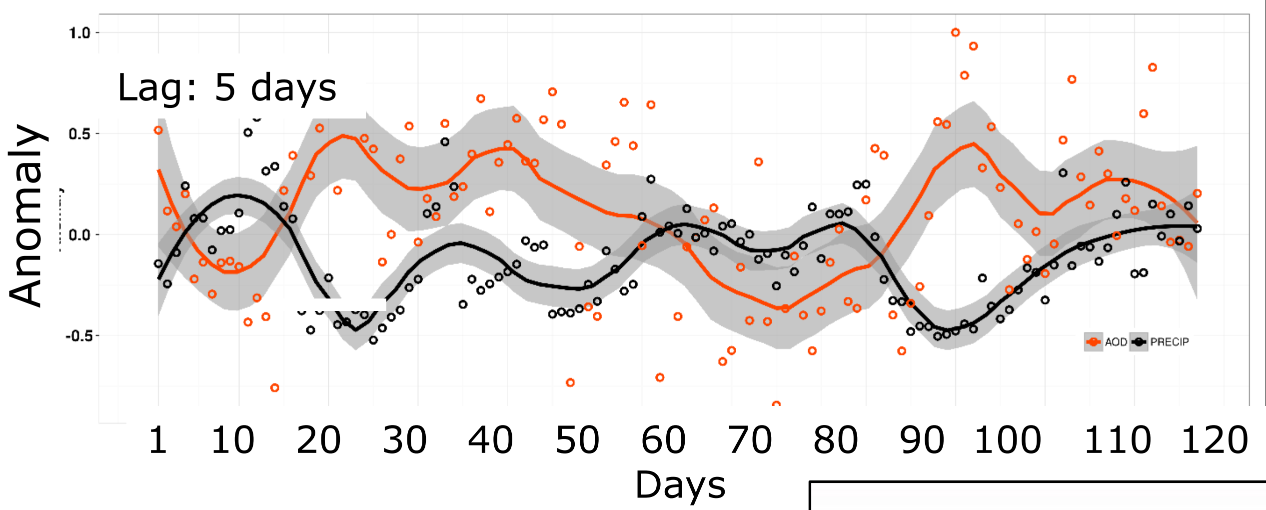
Increased frequency and decreased intensity of dry spells. Increased intensity of wet spells *(Singh et al., 2014).*

Both mean and wet extreme rainfall decreased over India and Northern China during 1979–2005 at a rate of 0.2%/decade *(Lin et al., 2018).*

Different spatial domains, statistical thresholds and definitions of wet and dry spells leads to contradictory results *(Goswami et al., 2006; Singh et al., 2014; Rajeevan et al., 2010; Lin et al., 2018).*

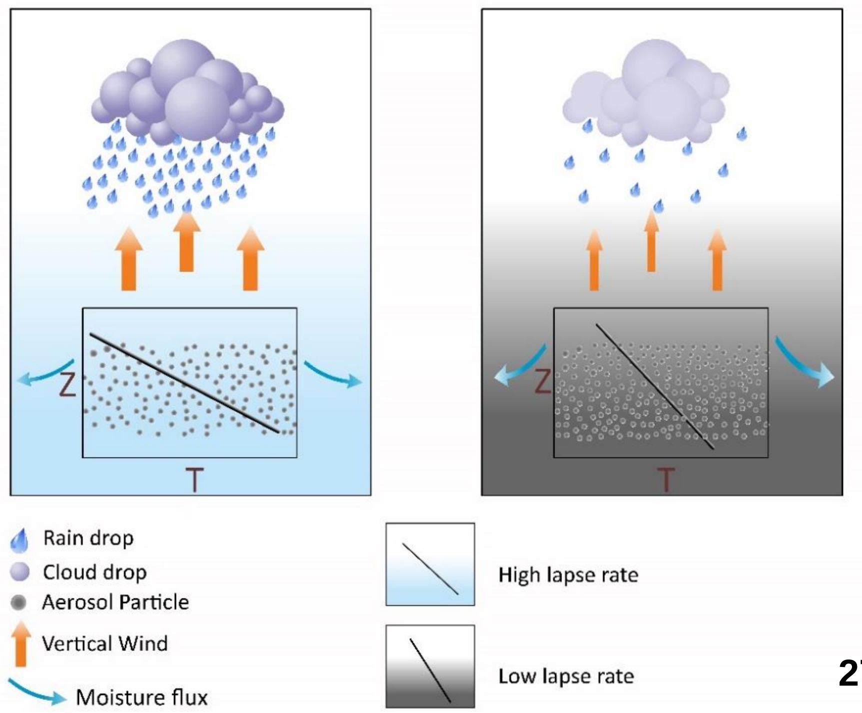
Spatial heterogeneity, differing response to extreme definition and complex response to forcing are characteristic of extreme rainfall analysis

Aerosol enhancement Granger causes intra-seasonal suppression of lagged daily rainfall (2000-2009)



- ❑ Correlation and causality analysis (2000-2010) to link aerosol enhancement (AOD, AAI) to rainfall, in regions of high/low aerosol & rainfall (HL/LL).

- ❑ AOD and AAI increases had causal influence on decreases in rainfall, with lag times of 1-5 days, in high aerosol – low rainfall regions, 3-4 times a season.
- ❑ This acts through atmospheric stabilization, increased divergence of moisture, along with reduced convection.
- ❑ A higher frequency of prolonged rainfall breaks, longer than seven days, occurred in high aerosol-low rainfall regions.



Dave, Bhushan, Venkataraman 2017, Scientific Reports.

Rainfall Extremes: Observational and model data analysis

Observational data analysis: Trends and **correlations**

- **Significance: Mann-Kendall** and **R-tests.**
- **Causality: Granger causality with time lags to isolate causal effects.**

Model simulations and analysis

Model: ECHAM6.3-HAM3.2

Horizontal Resolution: $1.87^{o} \times 1.87^{o}$

LSM: JSBACH; CP: Tiedtke scheme; SL: Monin–Obukhov theory; PBL: Mellor–Yamada scheme

AER: Hamburg Aerosol Module; CM: Lohmann scheme; RAD: PSRAD

DU Parameterization: Tegen et al., 2019; **India-specific dust tuning(in house)**

SS Parameterization: Long et al. (2011); Sofiev et al. (2011)

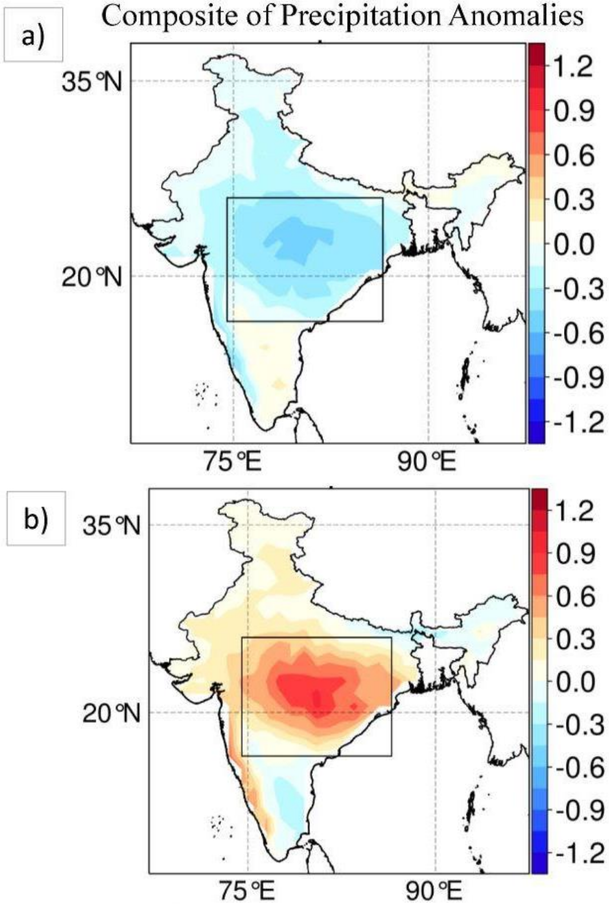
Model inputs

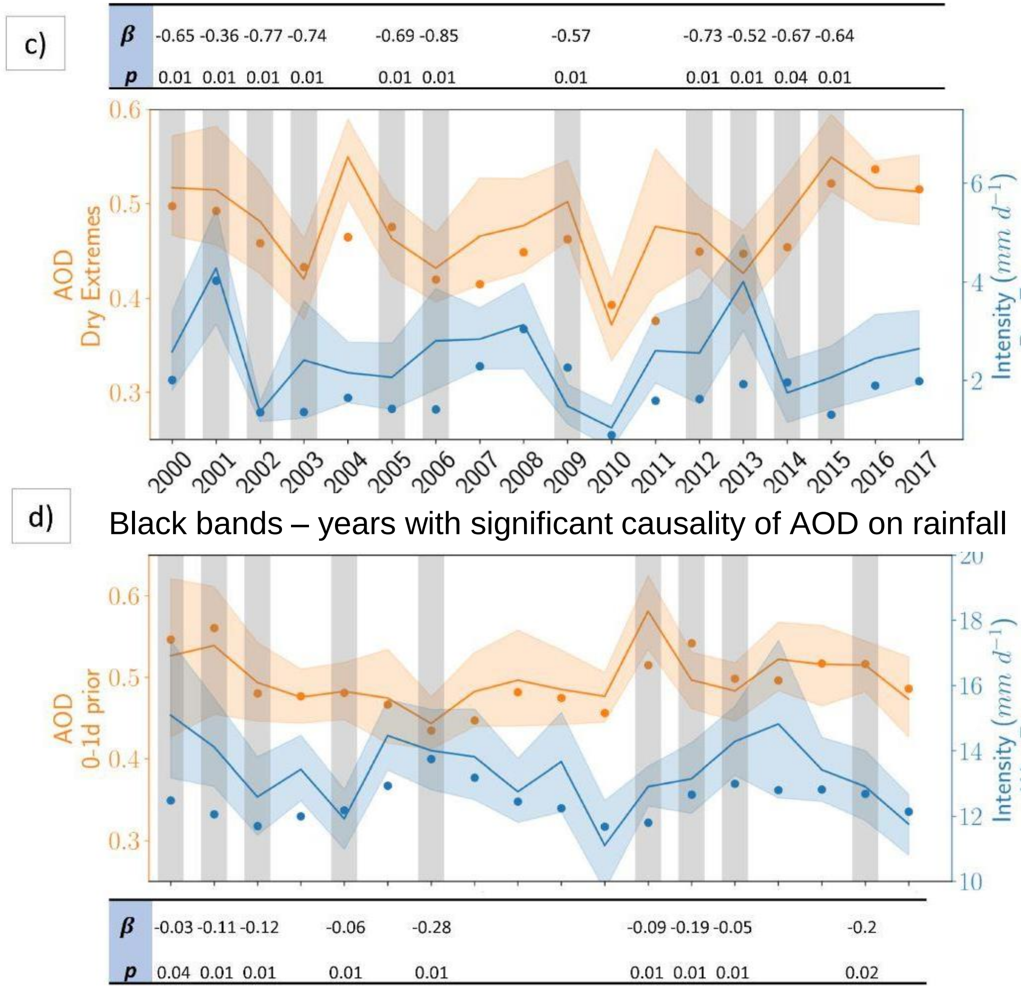
- ERA-I(Nudged data)
- HadISST, HadISIC
- Emissions(2005-2014)

Experiment design

HA: High aerosol or PD: Present Day from SMoG-India emissions nested in global inventory (CEDS).
PI: Pre-Industrial (1850) from (CEDS).
Composite vs climatology analysis.

Causal analysis of aerosol dessication of rainfall extremes (2001-2018)





$$rf_t = \alpha_0 + \sum_i^{lags} \alpha_i rf_{t-i} + \sum_i^{lags} \beta_i AOD_{t-i+1} + \varepsilon_t$$

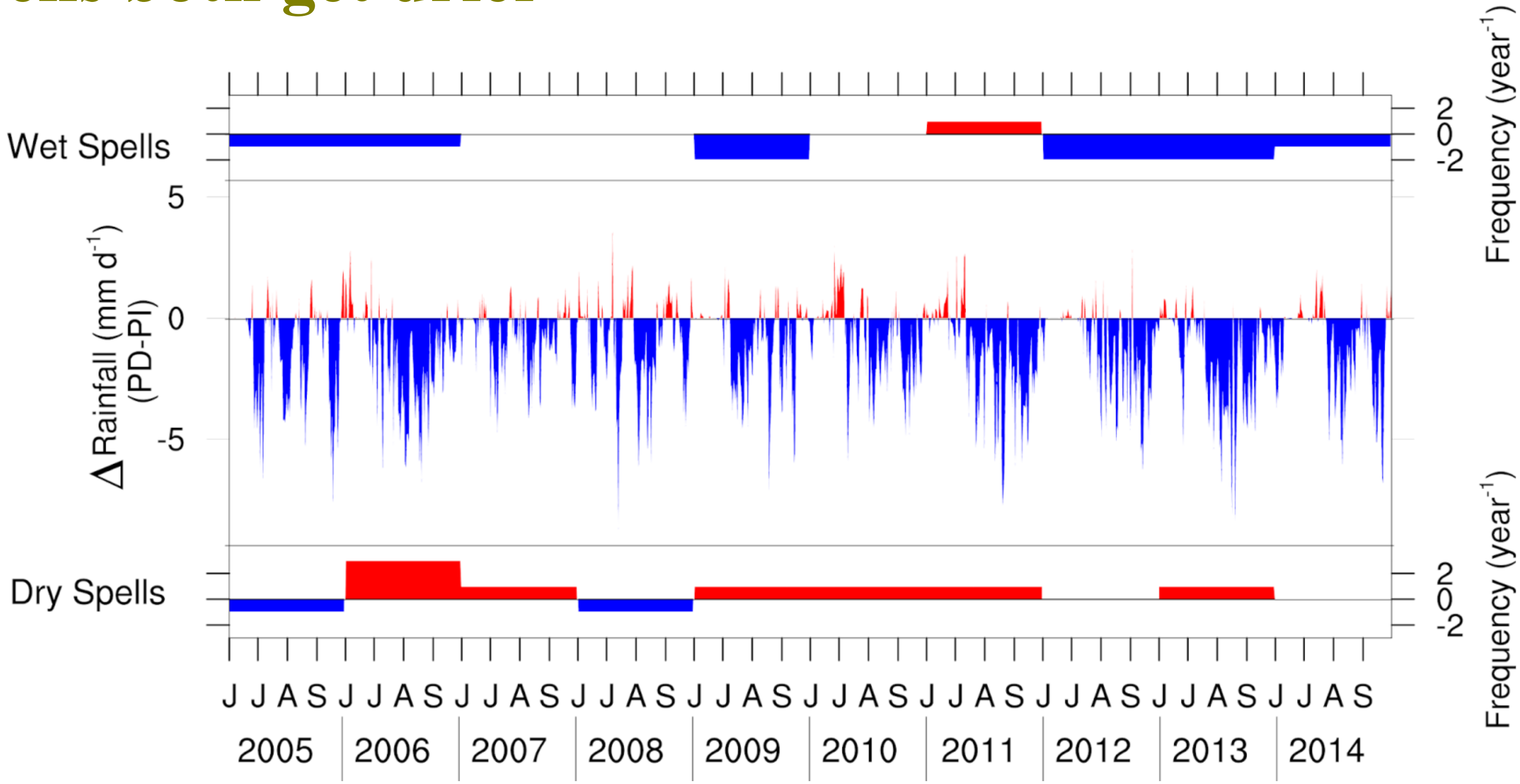
$$rf_t = \alpha_0 + \sum_i^{lags} \alpha_i rf_{t-i} + \varepsilon_t$$

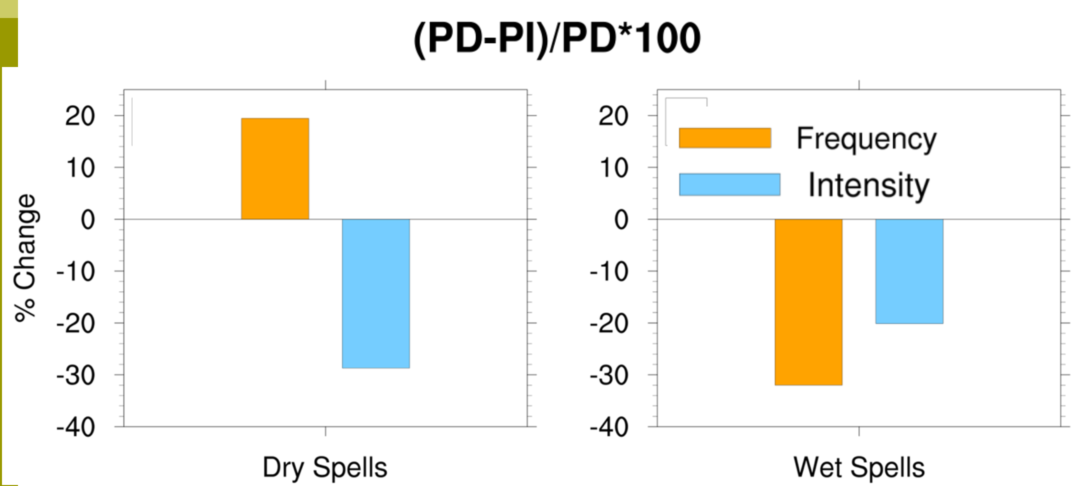
α-autoregression coefficient β- regression coefficient of rainfall with AOD
ε- residuals of the regression; p-tested for AOD causing rainfall having residuals lower than autoregression of rainfall.

- ❑ Causal effects of coincident aerosols over the "core" monsoon region: 60% of years show causality of AOD on rainfall suppression during the dry extremes and 40% for the wet extremes.
- ❑ Aerosol levels prior to wet spells used; reverse causality of rainfall to aerosol level ruled out.

Bhattacharya, Muduchuru, Venkataraman, Mondal, Das, 2022. submitted.

GCM simulations support the findings – wet and dry spells both get drier

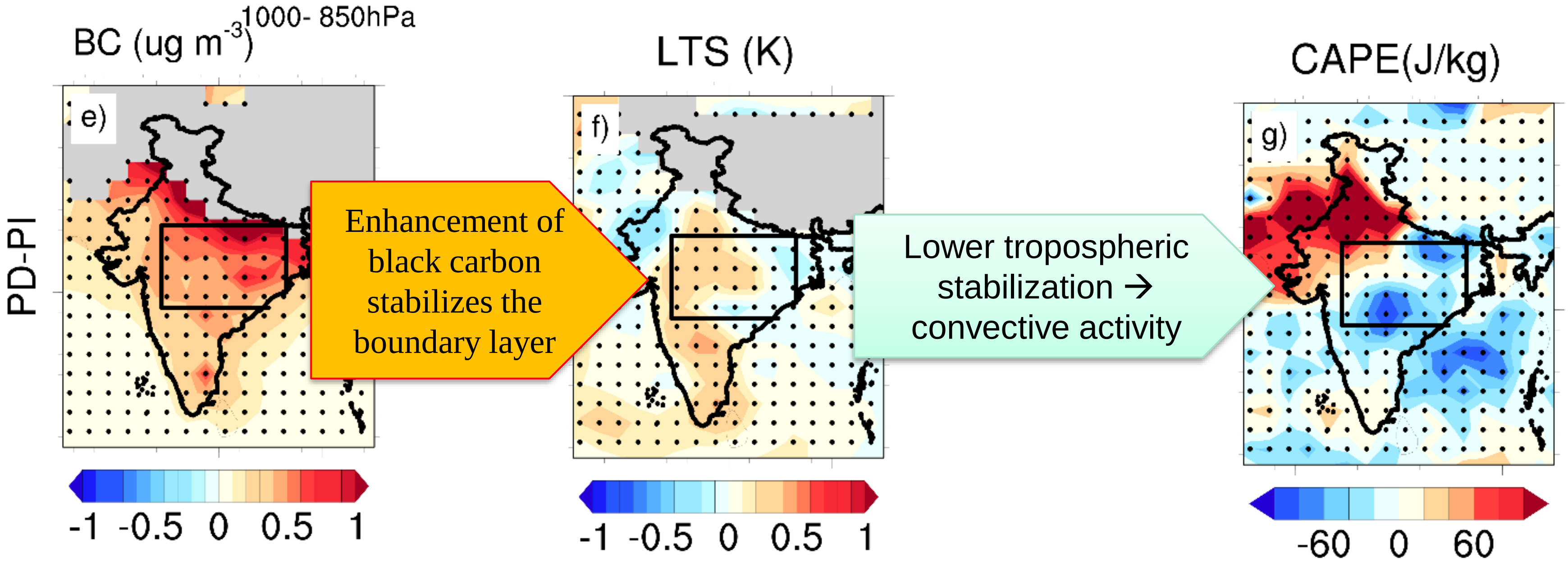




The model incorporates all direct effects of aerosols, the cloud albedo effect and the cloud lifetime effect.

Significance: Aerosol induced changes show that dry and wet spells both are getting drier.

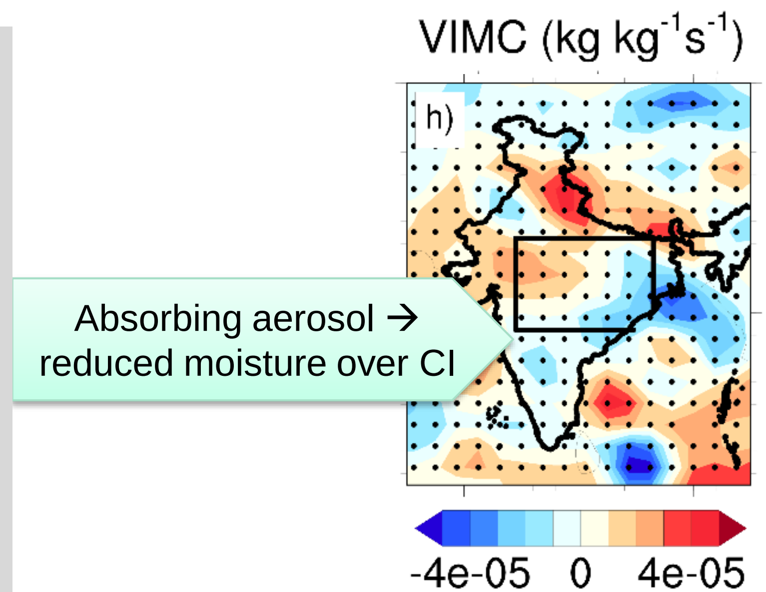
Black carbon radiative feedback underlies the drying of dry extremes



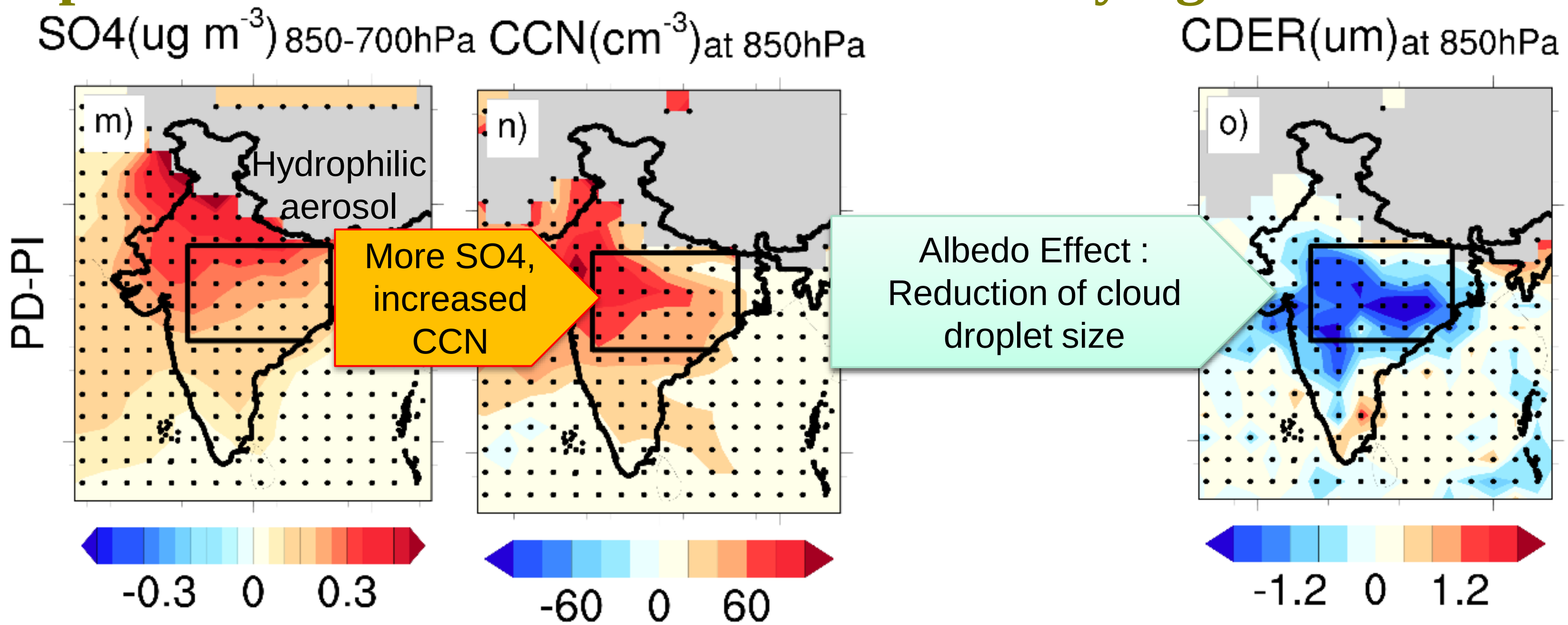
Dry extremes become more drier (intensity) through a aerosol induced **radiative** pathway.

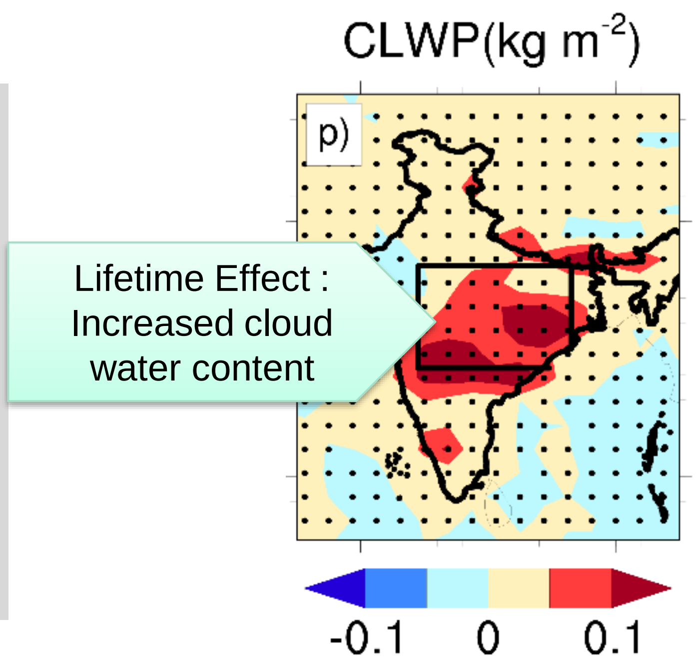
Black carbon differentially heats lower troposphere causing stabilization and lower convective activity, also reducing moisture over CI.

Cloud pathways did not show a discernible effect for dry extremes.



Sulphate-cloud interaction underlies the drying of wet extremes





Wet extremes become drier (intensity reduction) through a **aerosol cloud interaction** pathway.

Sulfate concentrations (@cloud layers) shown to increase CCN, reducing effective size of cloud droplets.

The reduction of cloud droplet size reduces rainfall efficiency during wet spells.

Perspective, uncertainties, open questions

- Land-atmosphere coupling processes affect feedbacks to temperature and water vapour; need improvements in turbulent heat transfer and surface layer vertical profiles, among others.
- Cloud convective and stratiform parameterizations need explicit links to aerosols.
- Anomalous regional cloud behaviour "anti-Twomey effect" cannot be explained by models.
- GCMs are able to capture both mean and extreme rainfall changes, through slow and fast response pathways.
- Stratiform processes indicate high spatial heterogeneity; need further evaluation of stratiform parameterizations; can only capture "Twomey effect".
- Cloud processes like invigoration, interstitial heating not parameterized in most models.

Thank you, questions?

Extra Slides

Aerosol influence on rainfall extremes: observational and GCM study

Observational studies to statistically establish aerosol role on rainfall extremes:

- Correlation-causality analysis on frequency, intensity and duration of rainfall extremes.
- Multi-annual trends of AOD and rainfall extreme indices (R10, Rx3,R1)

Atmospheric GCM simulations to understand mechanisms of aerosol influences on extreme rainfall:

- Quantitative response of rainfall extremes to present and pre-industrial aerosol levels over India.
- Mechanism through moist static energy and cloud microphysical changes.

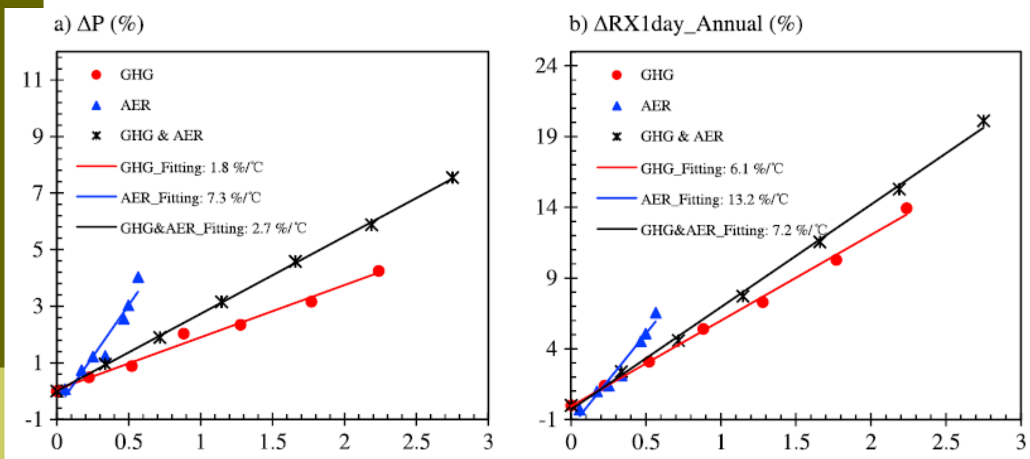
Math | Granger causality

$$X_t = \alpha_0 + \alpha_1 X_{t-1} + \ldots\ldots + \alpha_n X_{t-n} + \varepsilon_x$$

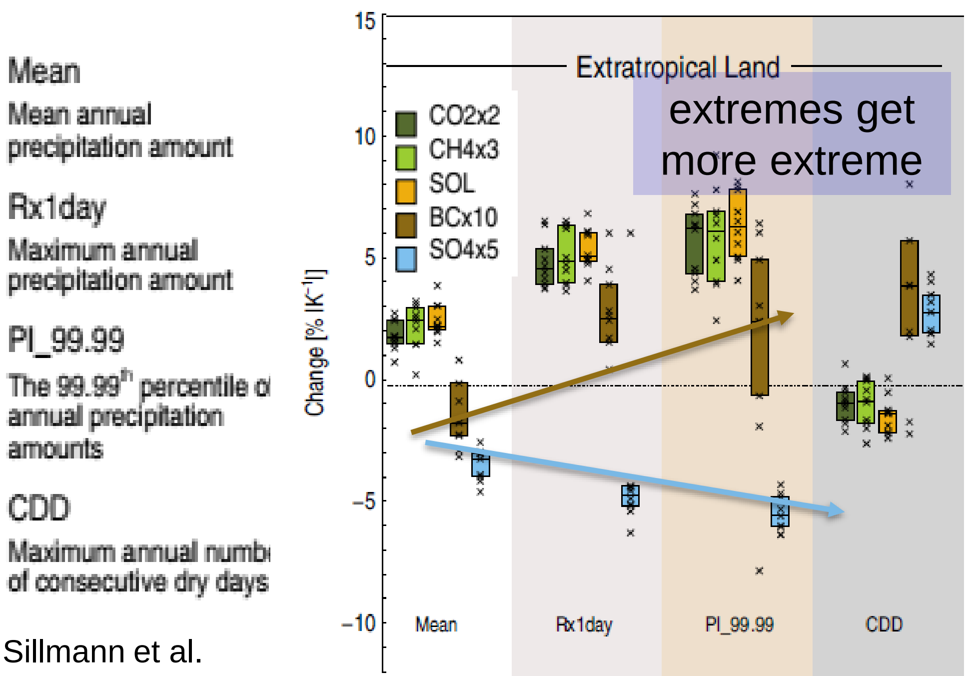
$$X_t = \alpha_0 + \alpha_1 X_{t-1} + \ldots\ldots + \alpha_n X_{t-n} + \beta_1 Y_{t-1} + \ldots\ldots + \beta_n Y_{t-n} + \varepsilon_{xy}$$

If the variance of the residual in the second model, labelled $\sigma^2_{\varepsilon_{xy}}$, is significantly less than the variance of the residual in the first model, labelled $\sigma^2_{\varepsilon_x}$, then the inclusion of information of Y is improving the prediction of X_t implying that Y is Granger causing X. *GC* was tested at varying

Extreme rainfall: Aerosol vs GHG forcing



Lin et al. 2016



Sillmann et al. 2019

Increased rate of precipitation extremes caused by aerosol forcing is significantly larger than that caused by GHG forcing *(Lin et al., 2016).*

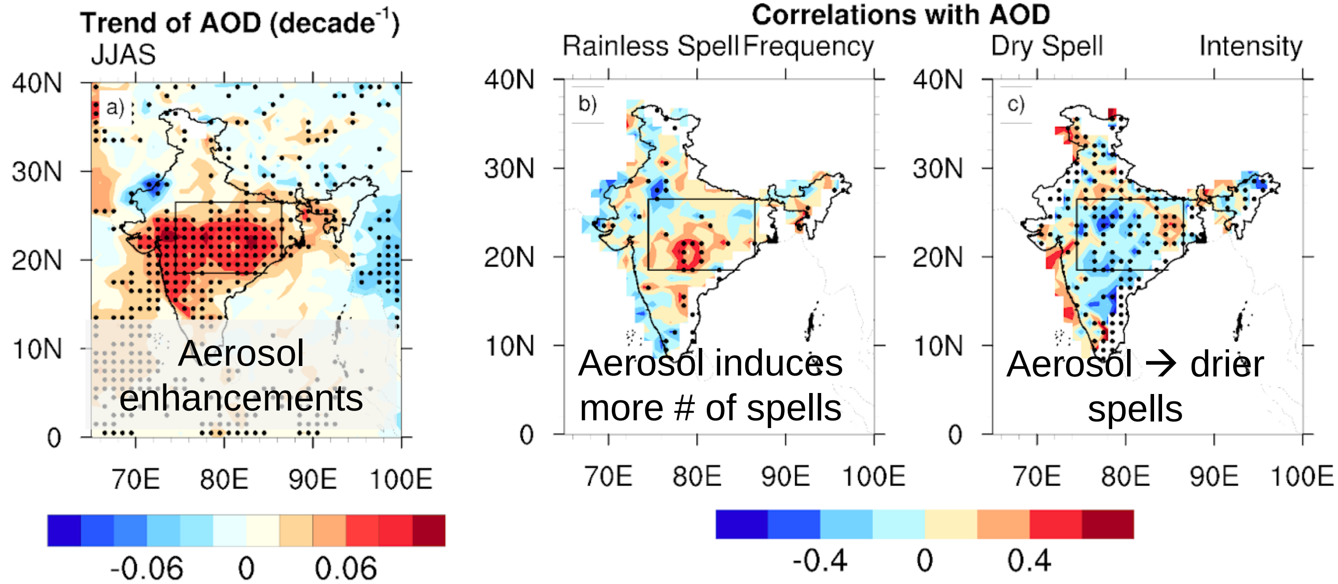
Globally, future RCP8.5, with reduced AOD compared to fixed 2005 levels increased intensity of wet extremes *(Zhao et al., 2019; Li et al., 2015).*

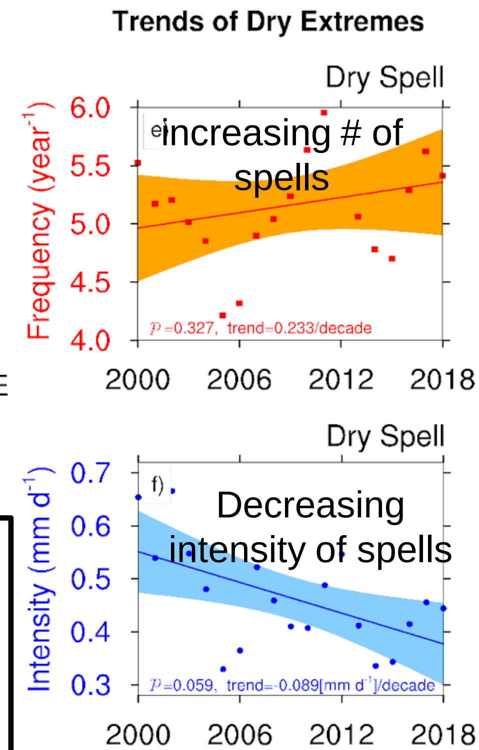
Dry days become more frequent with aerosols compared to GHGs

Absorbing Aerosols shown to increase wet and dry extremes (dry becomes drier and wet becomes wetter)

Scattering aerosols shown to decrease wet extreme frequency, but increase intensity *(Sillmann et al., 2019; Lin et al. 2018).*

Aerosol induced desiccation of the dry extremes





Definitions of different dry spells

Rainless Spell: Precipitation intensity less than 1 mm/day (Singh et al., 2019) for at least consecutive 3 days over a grid.

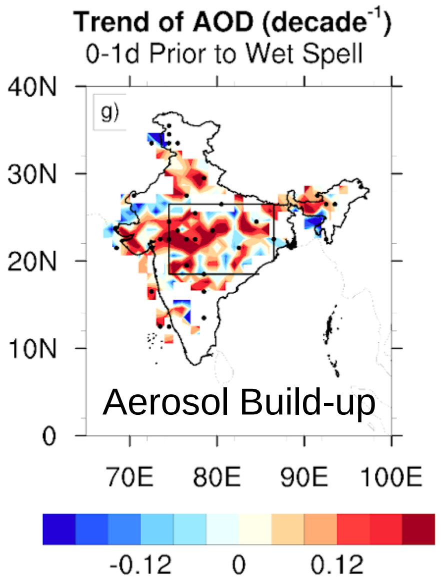
Dry Spell: Precipitation is less than (mean - 0.5stddev) for at least consecutive 3 days (Bhattacharya et al., 2017) over a grid.

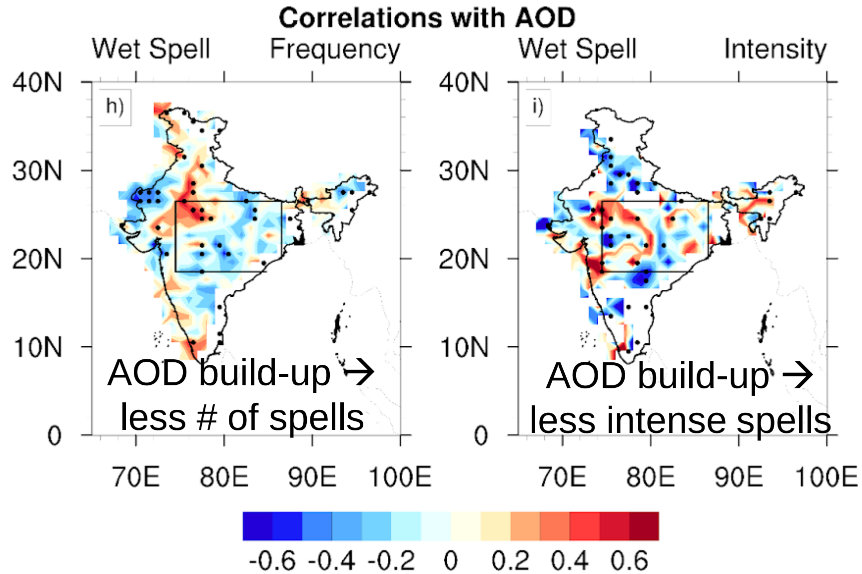
→ Intensity of dry spells decreases with year. Intensity negatively correlates with AOD enhancement.

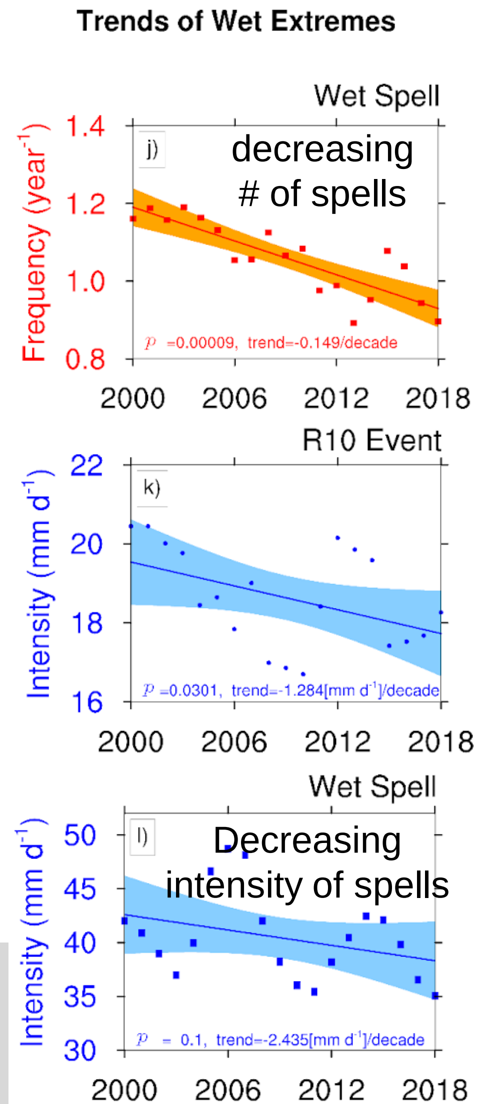
→Frequency increases over the recent period. Positive correlation between frequency of dry spells and AOD.

Aerosol increases are linked to a "dry gets drier" phenomenon (increasing frequency and decreasing intensity).

Aerosol induced drying of wet extremes







Definitions of different wet extremes

Wet Spell: Precipitation is greater than mean + 1 stddev for at least consecutive 3 days (Singh et al., 2014) over a grid.

R10 Event: Precipitation intensity is greater than 10 mm/day (Bhattacharya et al., 2017) over a grid.

Intensity of wet extremes decreases with time; negatively correlated with AOD build-up prior to the onset of the spells.

Wet spells/events is getting drier linked to aerosol enhancements.

Dry static energy balance at local scales

Energetic calculations (Muller and O'Gorman 2011) proposed to include dry static energy to account for local-scale effects. A dry static energy (DSE) flux divergence term is introduced to account for the local scale circulation changes.

$$L_c\delta P = \delta Q + \delta H = \delta LW + \delta SW - \delta SH + \delta H$$

Where δ → perturbation (HA-LA) due to aerosol changes
L_c → latent heat of condensation, P → precipitation
Q → atmospheric diabatic cooling = (atmospheric abs. (LW+SW) and sensible heat from surface)

Further δH is split into components arising from advective changes and thermodynamical changes

$$\delta H = \delta H_{Dyn_v} + \delta H_{Thermo_v} + \delta H_{Dyn_h} + \delta H_{Thermo_h}$$

$$= \int \delta(\bar{\omega})\frac{\partial \bar{s}}{\partial p} + \int \bar{\omega}\delta\left(\frac{\partial \bar{s}}{\partial p}\right) + \int \delta(\bar{\boldsymbol{u}}).\nabla\,\bar{s} + \int \bar{\boldsymbol{u}}.\delta(\nabla\,\bar{s})$$

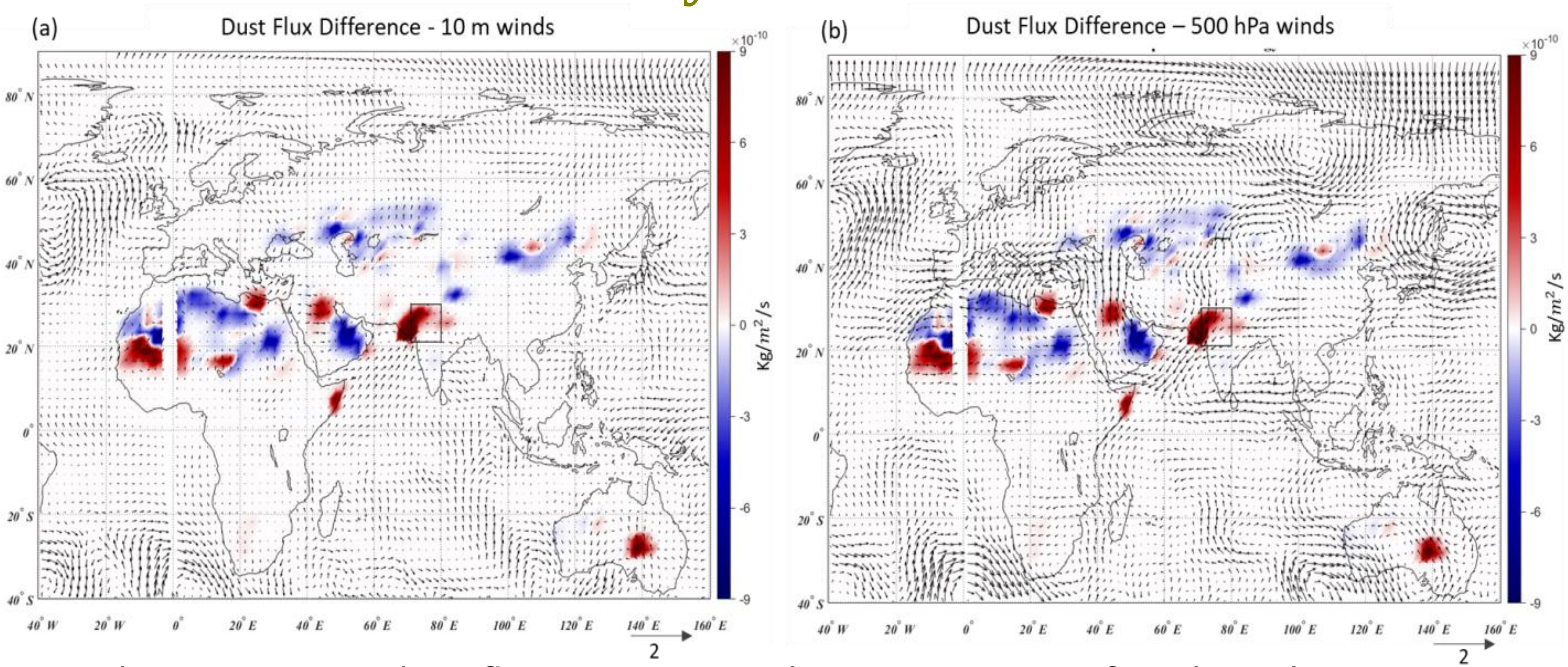
$$\nabla\bar{s} = \frac{1}{r}\frac{\partial \bar{s}}{\partial \theta}\hat{\theta} + \frac{1}{r\sin\theta}\frac{\partial \bar{s}}{\partial \varphi}\hat{\varphi}$$

$$s = gz + C_pT$$

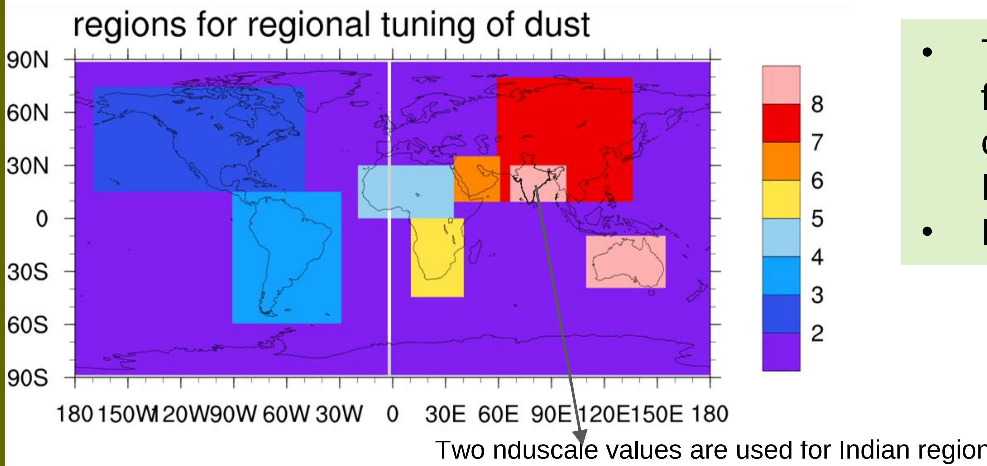
Where ω → vertical velocity, s → dry static energy, p → pressure, $\boldsymbol{u}$ → horizontal wind vector, z → model geopotential height, T → air temperature, Cp → isobaric specific heat of air, and the $\int$ denotes the mass-weighted vertical integration.

Overbars signify long term climatological averages. Here it is an average of 40 years of the model simulation time.

Meteorology affecting dust flux: threshold velocity linked to friction velocity

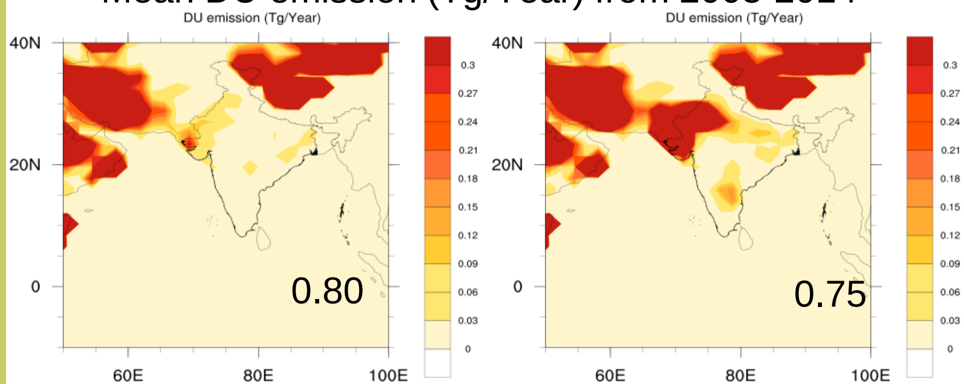


- Enhancement in dust flux over NW India, some parts of Arabia, the eastern horn of Africa and regions to the north of India.
- Difference in simulated u-v winds (at 10m height) shows an increase in surface wind magnitude near the lower end of the NW India box.
- Enhanced surface winds transport additional dust to NW India from the horn of Africa region.
- Higher level transport (u-v winds at 500 hPa), implies possible dust transport from the Arabian Peninsula and the Garagum and Taklamakan deserts to NW India.



- There is significant underestimation of emission flux in ECHAM6.3-HAM2.3 (nudged) as compared to ECHAM6.1-HAM2.2 (free) over Indian Landmass.
- It may be due to *"nudscale"* parameter.

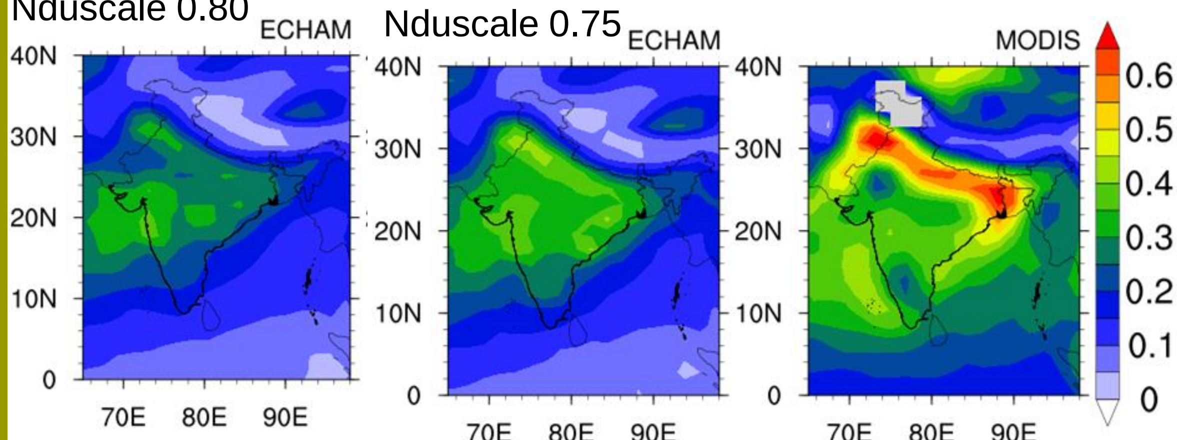
Mean DU emission (Tg/Year) from 2005-2014



Mean AOD from 2005-2014

Nduscale 0.80

Nduscale 0.75



Region	Simulation 1	Simulation 2
North America	1.25	1.25
South America	1.25	1.25
North Africa	0.95	0.95
South Africa	0.95	0.95
Middle East	0.95	0.95
Asia	1.25	1.25
India	0.8	0.75
Australia	0.95	0.95
Others	0.95	0.95

Dry deposition parameterization in ECHAM6-HAM2

Dust threshold friction velocity uncertain: Fluxes over Indian underestimated

Dry deposition flux (F_c) is a product of aerosol concentrations at a reference height (C_z) and deposition velocity (V_d) at the reference height z and is expressed as:

$$\boldsymbol{F_c = C_z V_d}$$

The deposition velocity $v_{d,k}$ of the kth moment is given by:

$$\boldsymbol{V_{d,k} = \frac{1}{r_a + r_{b,k} + r_a r_b V_{s,k}} + V_{s,k}}$$

Where r_a is the aerodynamic resistance, $r_{b,k}$ the quasi-laminar layer resistance and $V_{s,k}$ the sedimentation velocity ($\mathbf{V_{s,k}} = \frac{2}{9}\frac{r^2 \rho g Cc}{\mu}$; C_c = slip correction coefficient; μ is viscosity of air)

r_a is calculated from the roughness length z_0 and the boundary layer stability as below:

$$\boldsymbol{r_a = \frac{1}{u_* k}\left[ln\left(\frac{z}{Z_0}\right) - \emptyset\left(\frac{z}{L}\right)\right]}$$

Where, u_* is the friction velocity, κ the von-Karmann constant of 0.4, z is the reference height (i.e., half of the lowest model layer height), Φ is a dimensionless stability term, and L the Monin-Obukhov-length.

$r_{b,k}$ is parameterized as following:

$$\boldsymbol{r_{b,k} = \frac{1}{\epsilon_0 u_* E_T}}$$

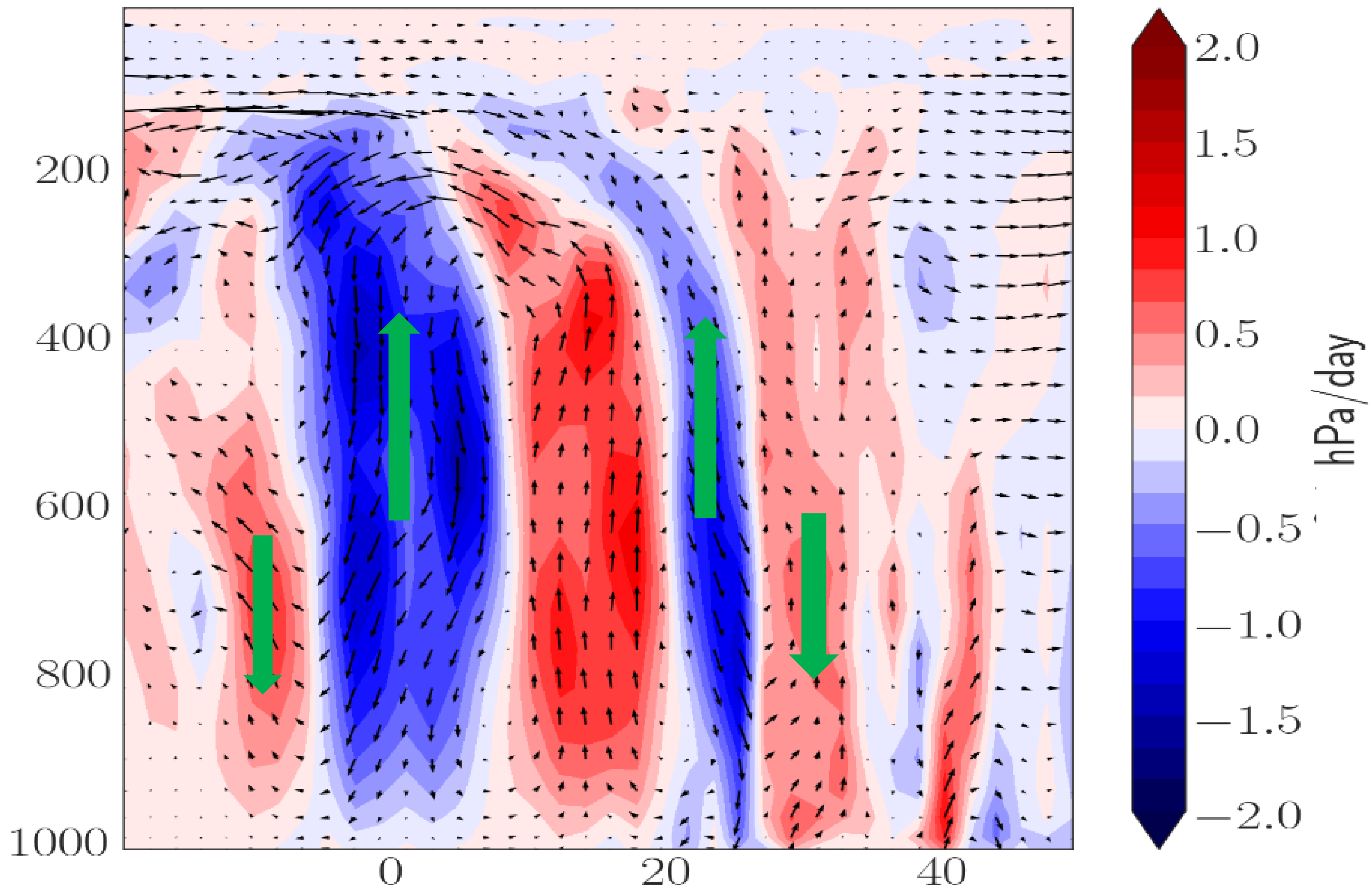
where E_T is the total collection efficiency and is parameterized as a series of collection efficiencies for each particle-surface interaction processes such as Brownian diffusion (E_B), interception (E_{IN})and impaction (E_{IM}) collection efficiency with the collecting surface i.e., $\mathbf{E_T = R\,(E_B + E_{IM} + E_{IN})}$

$\mathbf{R} = \boldsymbol{e^{-St^{1/2}}}$; $\mathbf{E_B = Sc^{-\gamma}}$; $\mathbf{E_{IM}} = \boldsymbol{\frac{St^2}{1+St^2}}$; $\mathbf{E_{IN}} = \boldsymbol{\frac{1}{2}\left(\frac{D_p}{A}\right)^2}$

where, R= correction factor; St = stokes number; Sc = schmidt number; A = characteristic radius of collectors, depends on land use category

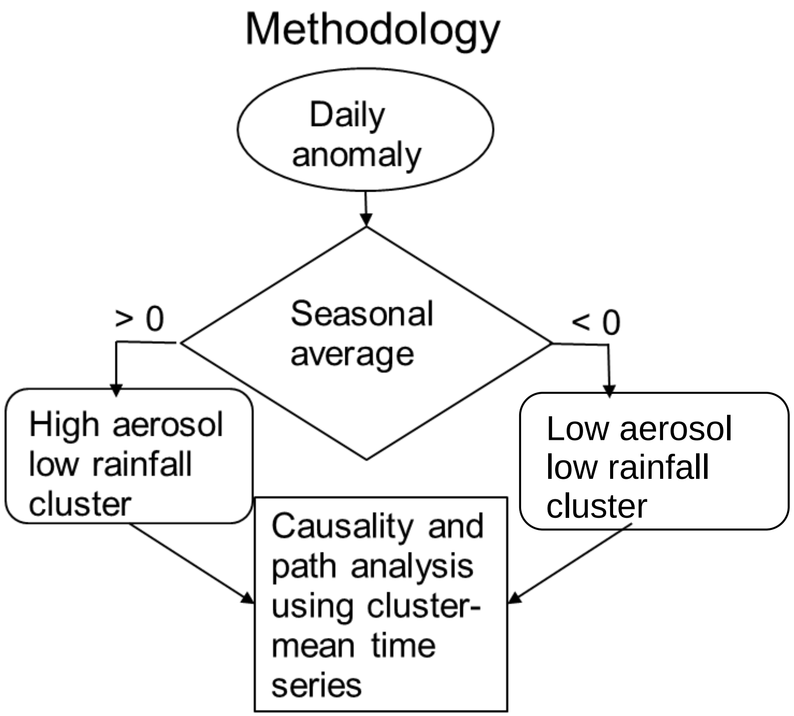
Hadley circulation response to aerosol emissions: July-Aug

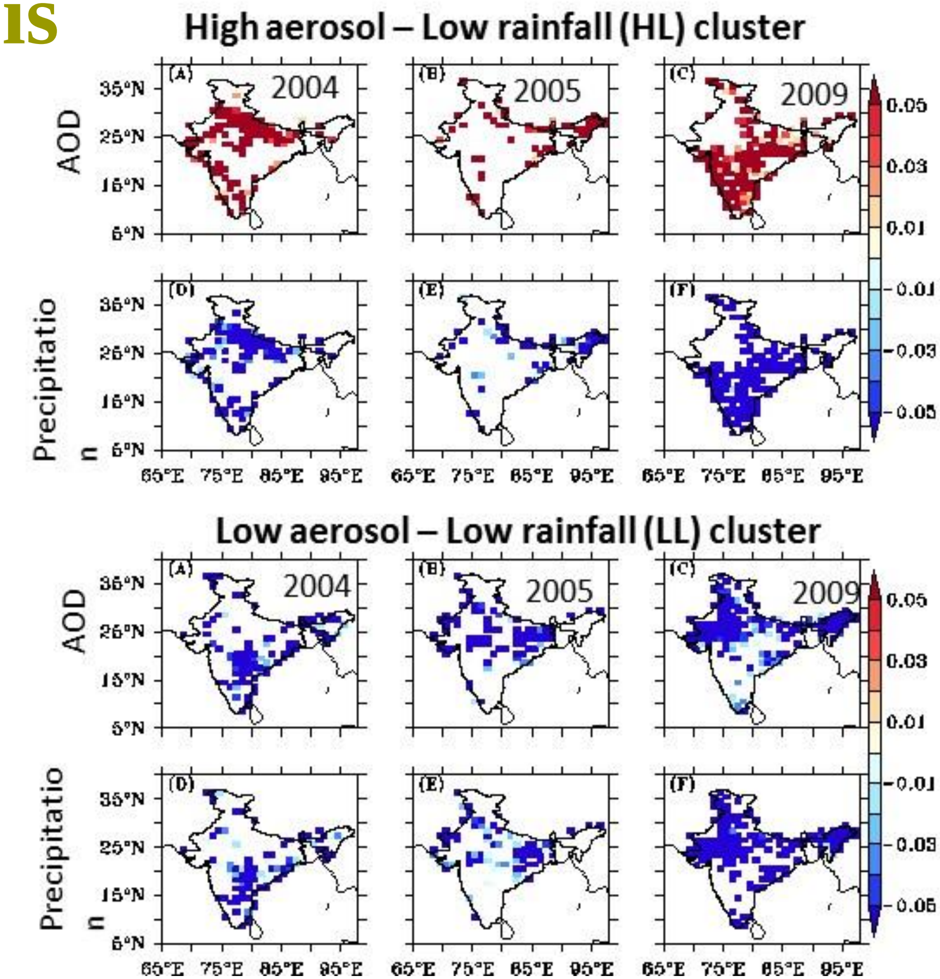
Aerosol response (H_{aero} minus L_{aero})



- Aerosol forcing breaks up the Hadley cell and weakens sinking / rising arms further south / north respectively, but with a persisting Hadley circulation between 0-20 N.

Aerosol causality analysis





Causality and path analysis:

Granger causality: If variable X increase prediction of variable Y, Y Granger causes X.
Path analysis: Overall effects segregated using path analysis.

- Years 2002, 2003, 2004, 2005, 2007 and 2009 showed regions extending over India, falling into HL and LL clusters.
- Years 2000 and 2001 did not have enough pixels (< 50); 2002 no influence of aerosols found, meteorological monsoon year (IMD, 2002).
- Therefore, years 2003, 2004, 2005, 2007 and 2009 were retained for further analysis of behavior in HL and LL clusters.

Dave et al. 2017, Scientific Reports