Observational constraints on aerosol indirect effects and controlling processes

Robert Wood
Atmospheric Sciences, University of Washington

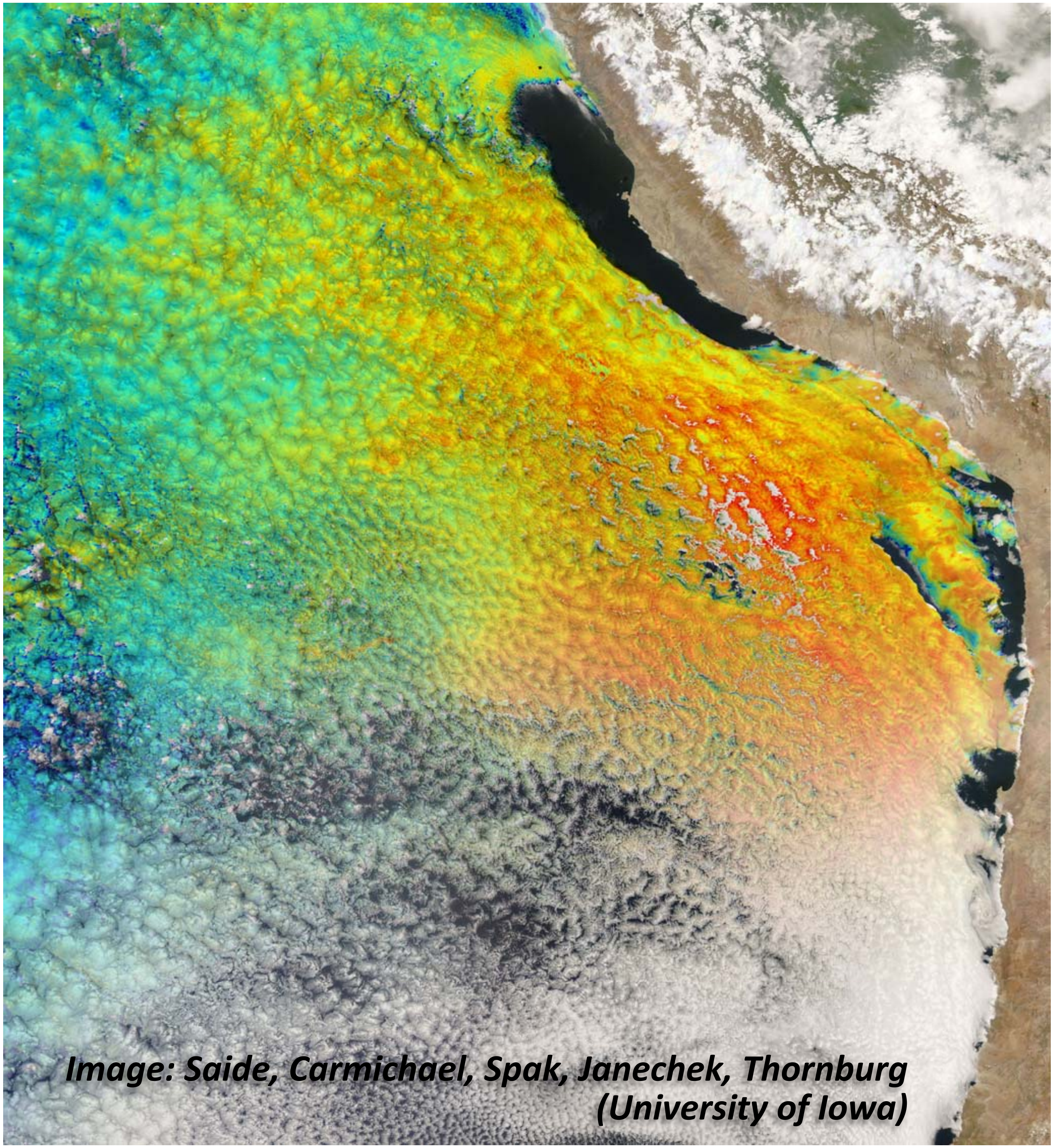
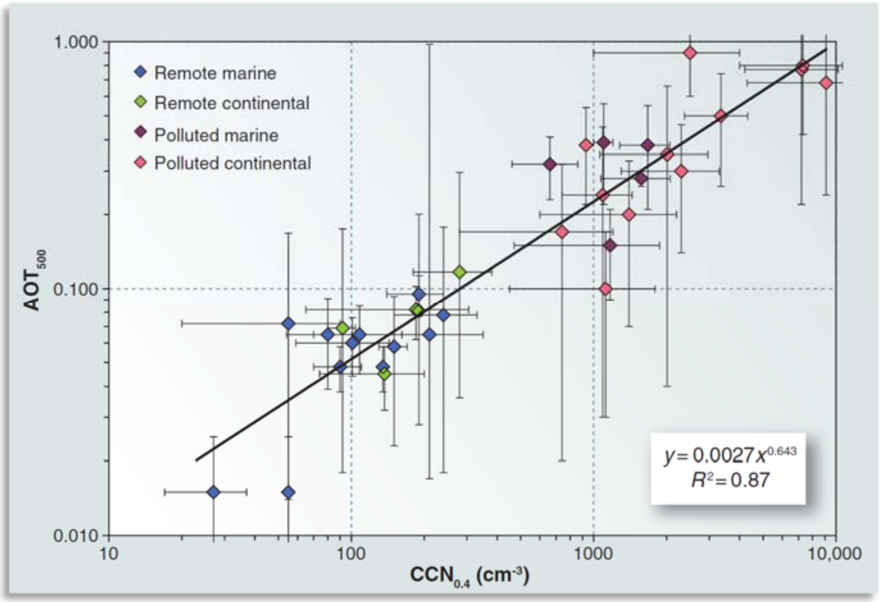


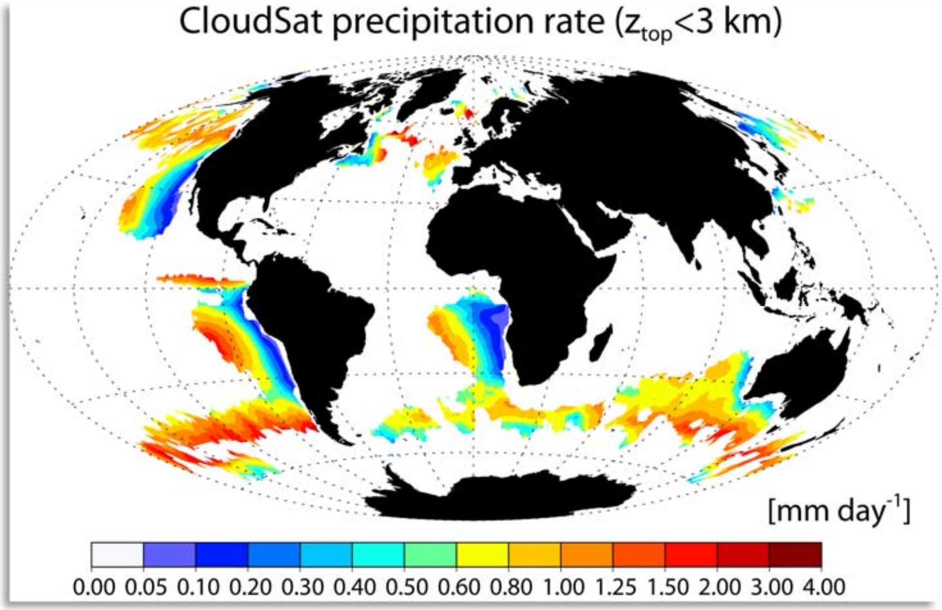
Image: Saide, Carmichael, Spak, Janechek, Thornburg (University of Iowa)

Motivating questions

- What are spaceborne aerosol measurements telling us about CCN?
- What are CCN over the remote oceans telling us about (anthropogenic aerosol sources)?



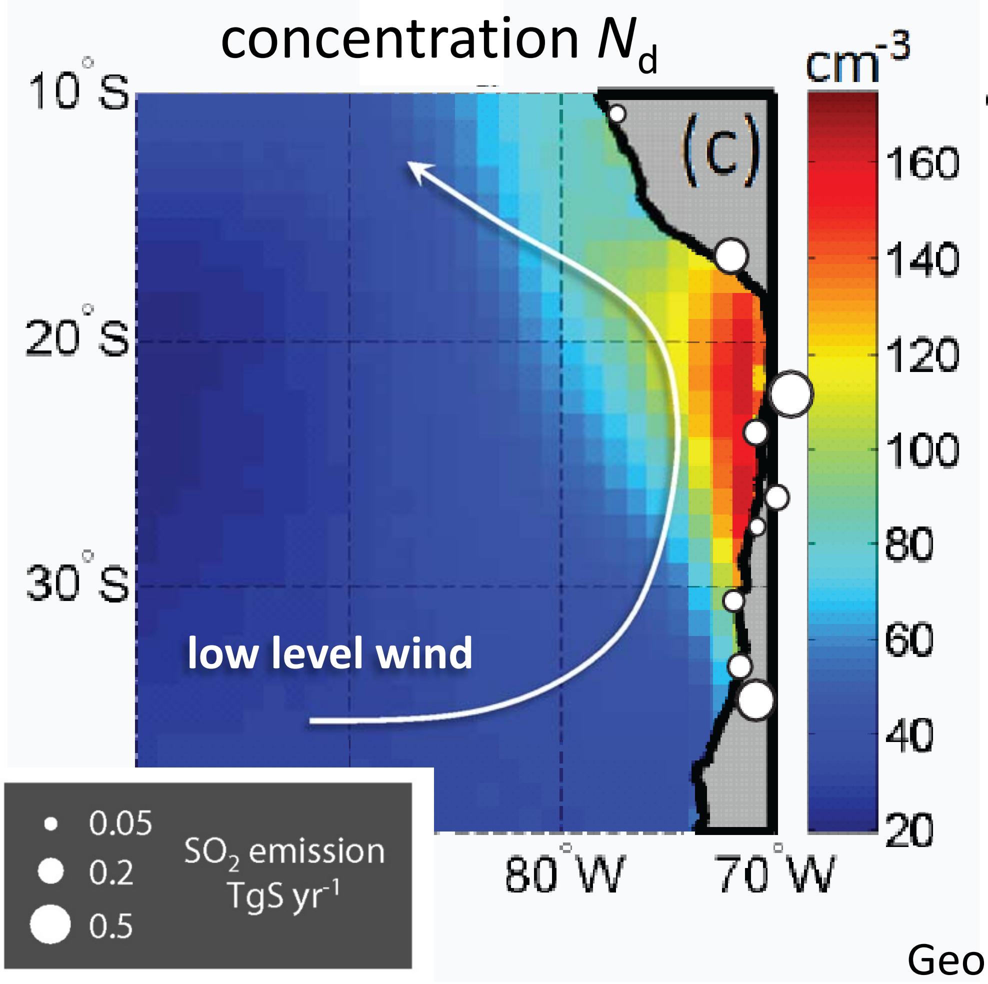
Rosenfeld et al., *Science*, 2008



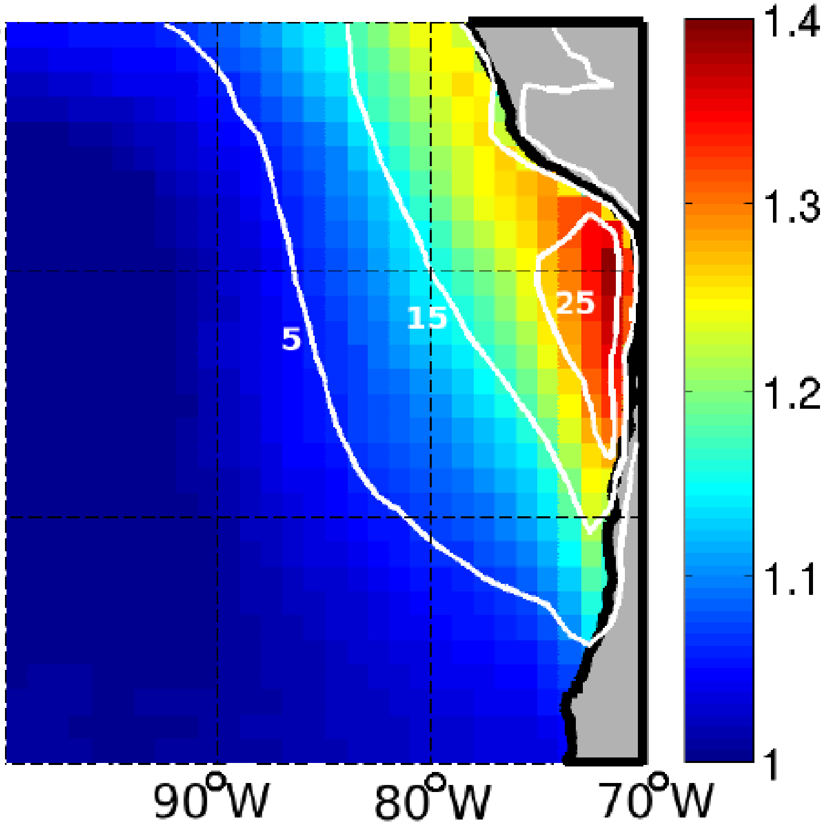
Wood et al *J. Geophys. Res.*, 2012

Regional gradients: Strong aerosol indirect effects in an extremely clean background

Satellite-derived cloud droplet concentration N_d



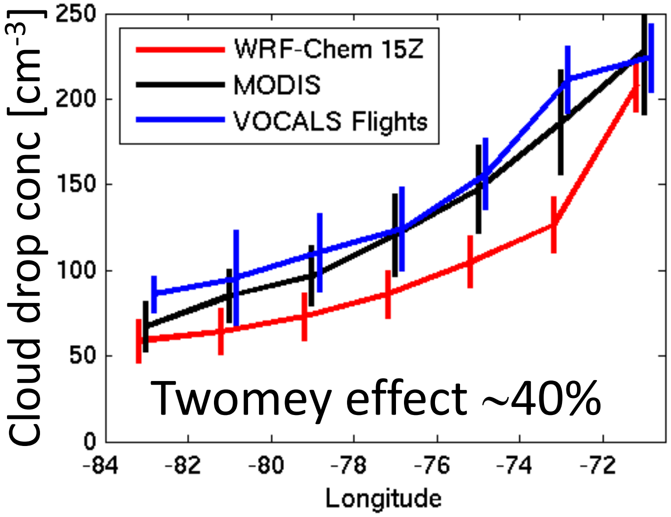
albedo enhancement (fractional)



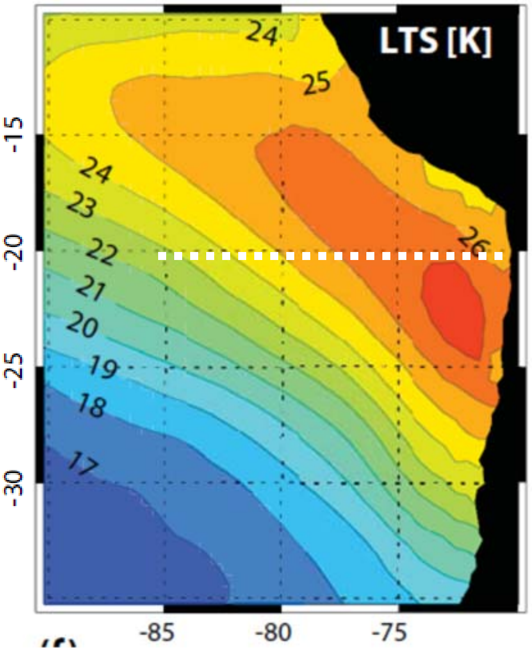
George and Wood, *Atmos. Chem. Phys.*, 2010

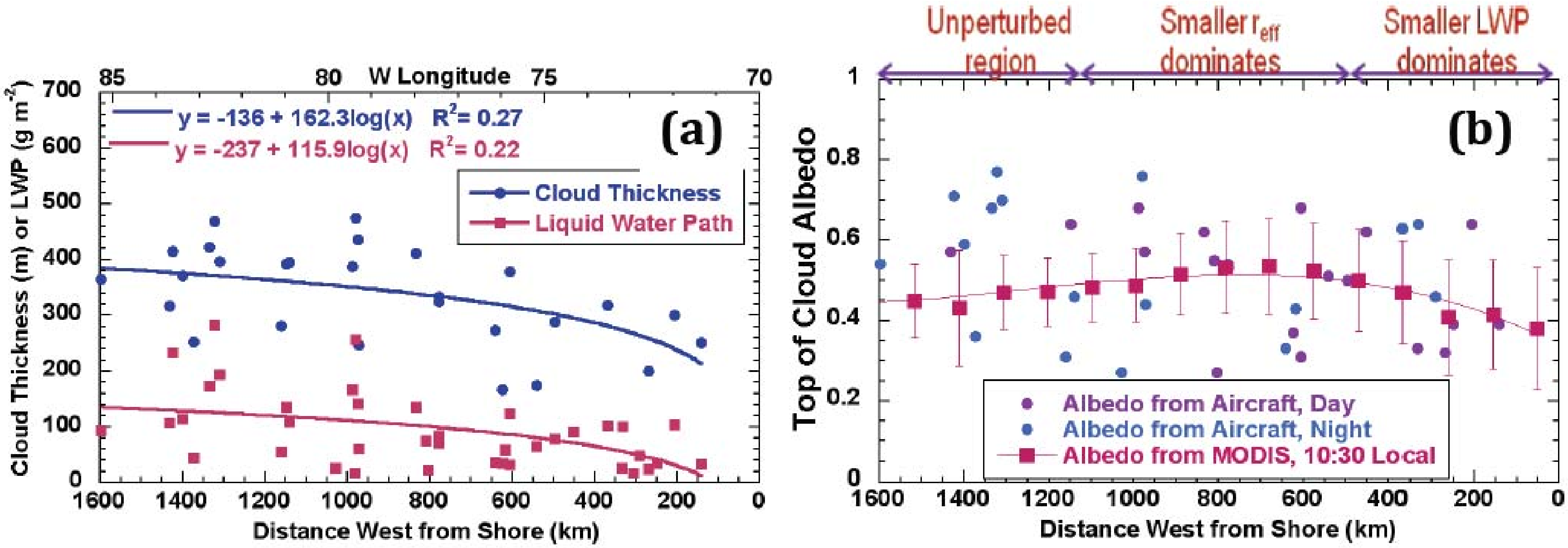
Regional gradients

- Cloud albedo changes dominated by meteorological effects despite large Twomey effect



Twomey effect ~40%

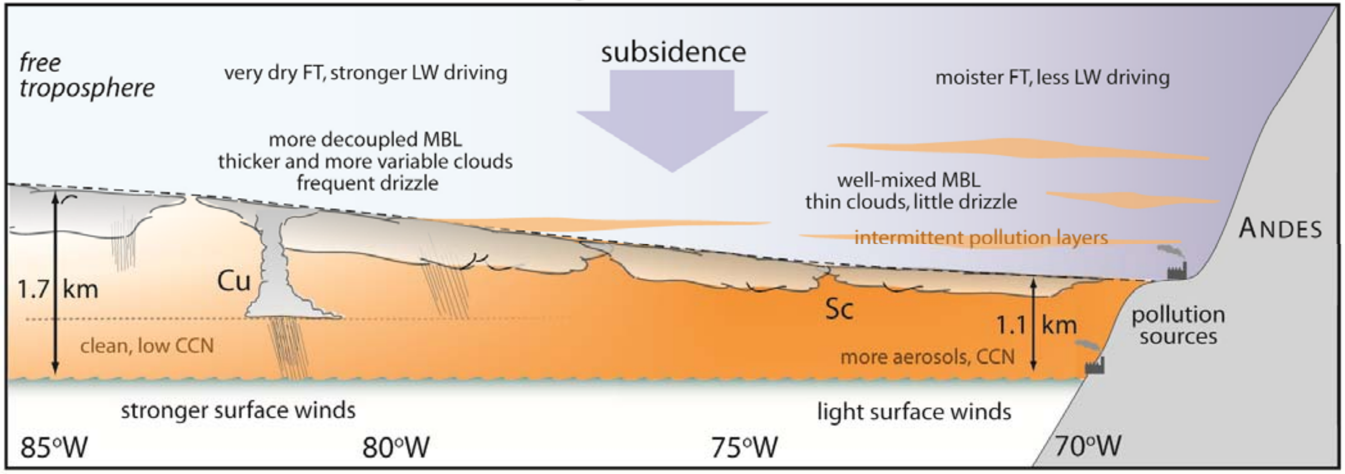




Twohy et al. (ACPD, 2012)

Does satellite AOD inform about CCN?

VOCALS as a testbed for understanding aerosol variability



- Break down aerosol optical depth τ into constituent parts

dry extinction

aerosol layer depth

$$\tau = \sigma G_\sigma h$$

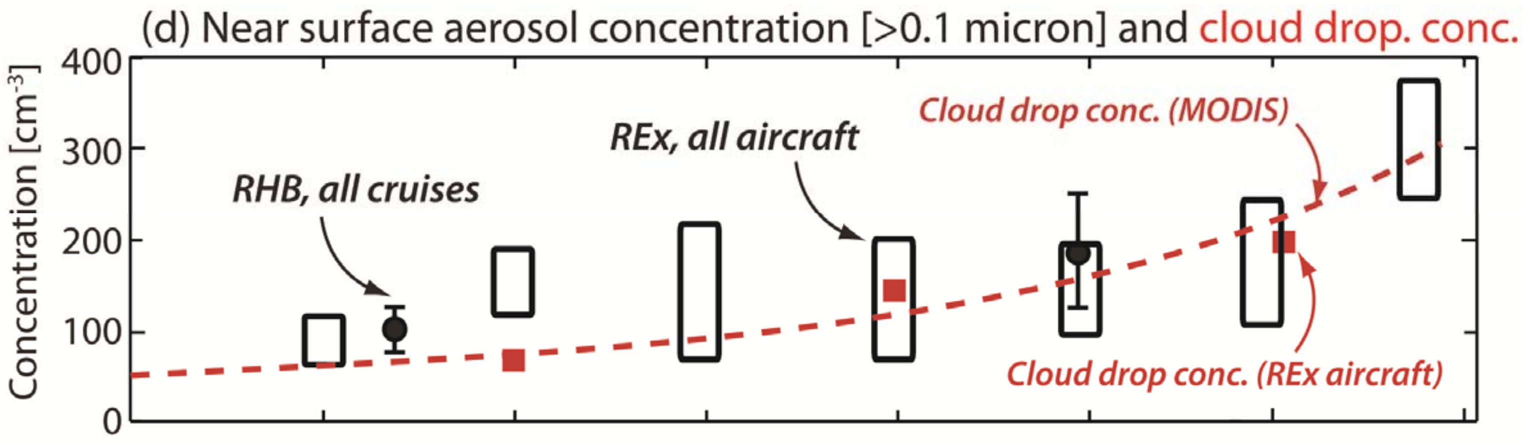
Hygroscopic growth

$$\frac{d\tau}{\tau} = \frac{d\sigma}{\sigma} + \frac{dG_\sigma}{G_\sigma} + \frac{dh}{h} = \frac{dN_a}{N_a} + 3\frac{dD_3}{D_3} + \frac{dG_\sigma}{G_\sigma} + \frac{dh}{h}$$

$$\frac{d\tau}{\tau} = \frac{d\sigma}{\sigma} + \frac{dG_\sigma}{G_\sigma} + \frac{dh}{h} = \frac{dN_a}{N_a} + 3\frac{dD_3}{D_3} + \frac{dG_\sigma}{G_\sigma} + \frac{dh}{h}$$

Three longitude bins: 80-85°W, 75-80°W, 70-75°W

dX is the increase from offshore to coastal bin; *X* is the mean value



Bretherton et al. (2010); Allen et al. (2011)

$$\frac{dN_a}{N_a} = \mathbf{0.72} \qquad \frac{dN_d}{N_d} = \mathbf{0.80}$$

$$\frac{d\tau}{\tau} = \frac{d\sigma}{\sigma} + \frac{dG_\sigma}{G_\sigma} + \frac{dh}{h} = \frac{dN_a}{N_a} + 3\frac{dD_3}{D_3} + \frac{dG_\sigma}{G_\sigma} + \frac{dh}{h}$$

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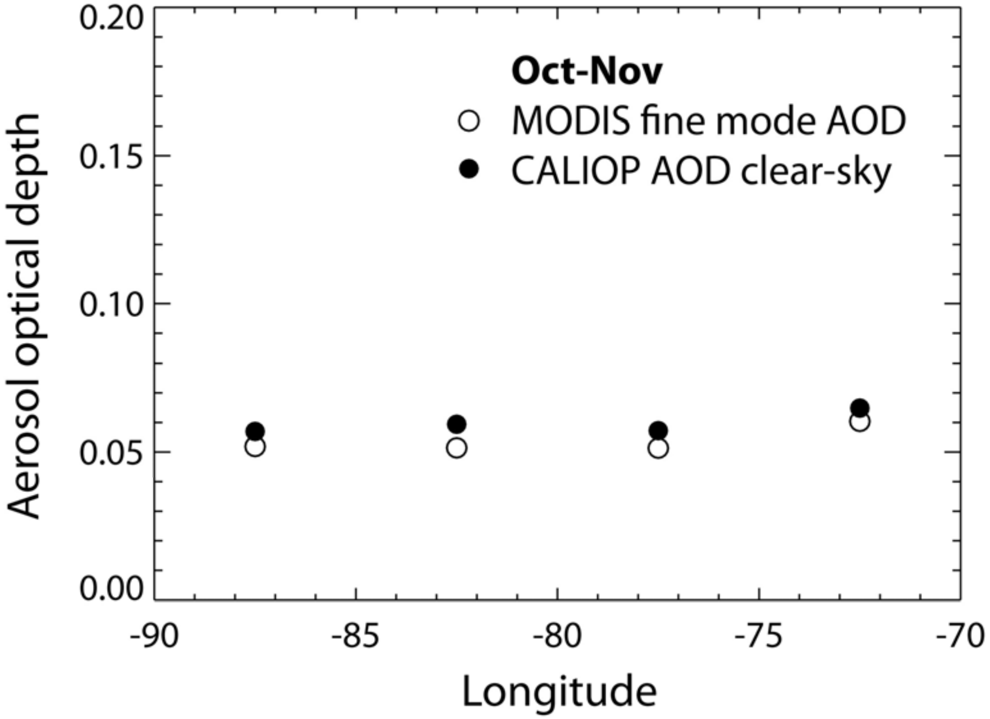

$$\frac{d\tau}{\tau} =$$ **0.09** (CALIOP)

0.17 (MODIS)

$$\frac{d\tau}{\tau} = \frac{d\sigma}{\sigma} + \frac{dG_\sigma}{G_\sigma} + \frac{dh}{h} = \frac{dN_a}{N_a} + 3\frac{dD_3}{D_3} + \frac{dG_\sigma}{G_\sigma} + \frac{dh}{h}$$

Three longitude bins: 80-85°W, 75-80°W, 70-75°W

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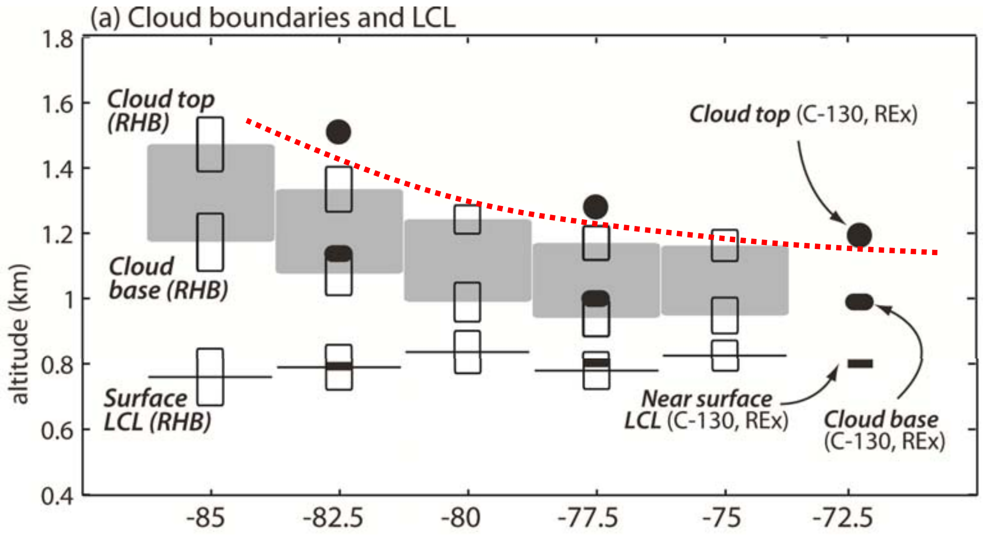
$$3\frac{dD_3}{D_3} = \mathbf{-0.38}$$ [D_3 decreases from 0.28 to 0.25 μm]

$$\frac{dh}{h} = \mathbf{-0.25}$$ [MBL depth decreases from 1.5 to 1.2 km]



Bretherton et al. (2010); de Szoeke et al. (2012)

$$\frac{d\tau}{\tau} = \frac{d\sigma}{\sigma} + \frac{dG_\sigma}{G_\sigma} + \frac{dh}{h} = \frac{dN_a}{N_a} + 3\frac{dD_3}{D_3} + \frac{dG_\sigma}{G_\sigma} + \frac{dh}{h}$$

$$\frac{d\sigma}{\sigma} = \mathbf{0.33} = \mathbf{0.72 - 0.38}$$

$$= 0.34$$

Nephelometer

PCASP

Closure achieved for dry aerosol scattering

$$\frac{d\tau}{\tau} = \frac{dN_a}{N_a} + 3\frac{dD_3}{D_3} + \frac{dG_\sigma}{G_\sigma} + \frac{dh}{h}$$

↓ ↓

$\frac{d\tau}{\tau}$ = **0.09** to **0.17**

$\frac{dN_a}{N_a}$ = **0.72**

$3\frac{dD_3}{D_3}$ = **–0.38**

$\frac{dh}{h}$ = **–0.25**

$\frac{dG_\sigma}{G_\sigma}$ = **0.0-0.24***

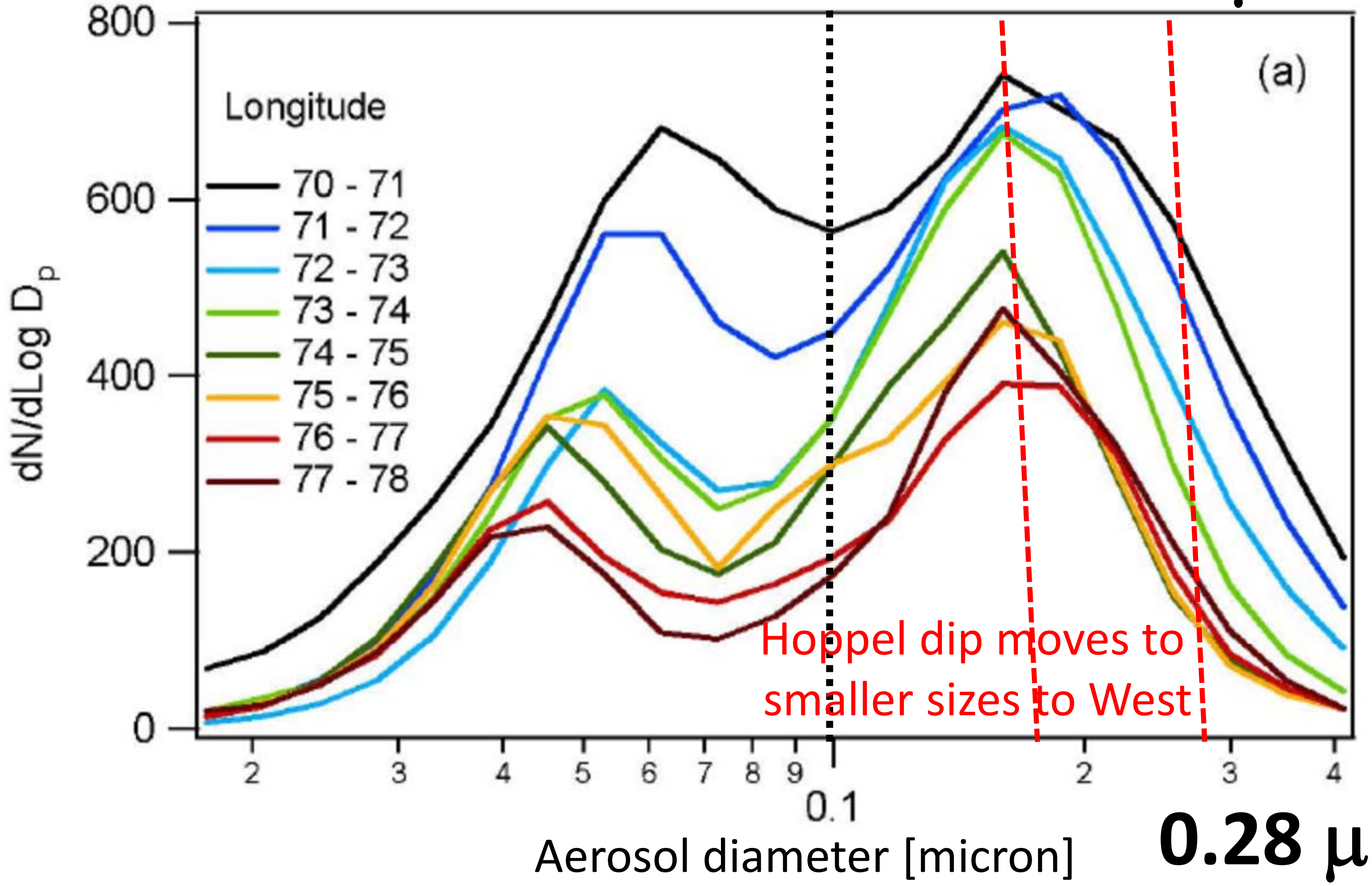
= **0.09-0.17** **0.09-0.33**

*highly uncertain

⇒ Large increase in conc. offset by reduction in size, and MBL depth

Changes in size are small, but important

0.25 μm



Hoppel dip moves to smaller sizes to West

0.28 μm

"The droplet mode size is nearly invariant" Kleinman et al. (ACP, 2011)

What controls CCN and cloud microphysical variability in the marine boundary layer?
A simple CCN budget for the PBL

$$\dot{N} = \overset{\text{Entrainment}}{\dot{N}_{\mathrm{FT}}} + \underset{\text{Surf. Source}}{\dot{N}_{\mathrm{S}}} + \overset{\text{Nucleation/Secondary}}{\dot{N}_{\mathrm{PROD}}} + \underset{\text{Precip. Sink}}{\dot{N}_{\mathrm{P}}} + \overset{\text{Dry Dep.}}{\dot{N}_{\mathrm{DRY}}} + \underset{\text{Advection}}{\dot{N}_{\mathrm{ADV}}}$$

- Assume nucleation/secondary processes unimportant
- Dry deposition is negligible (Georgi 1990)
- Sea-spray formulation (e.g. Clarke et al. 2006)
- Ignore advection
- Precipitation sink primarily from accretion process
- Equivalency of CCN and cloud drop conc. N_d

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Steady-state CCN budget

Free Tropospheric CCN

Sea-Spray Production

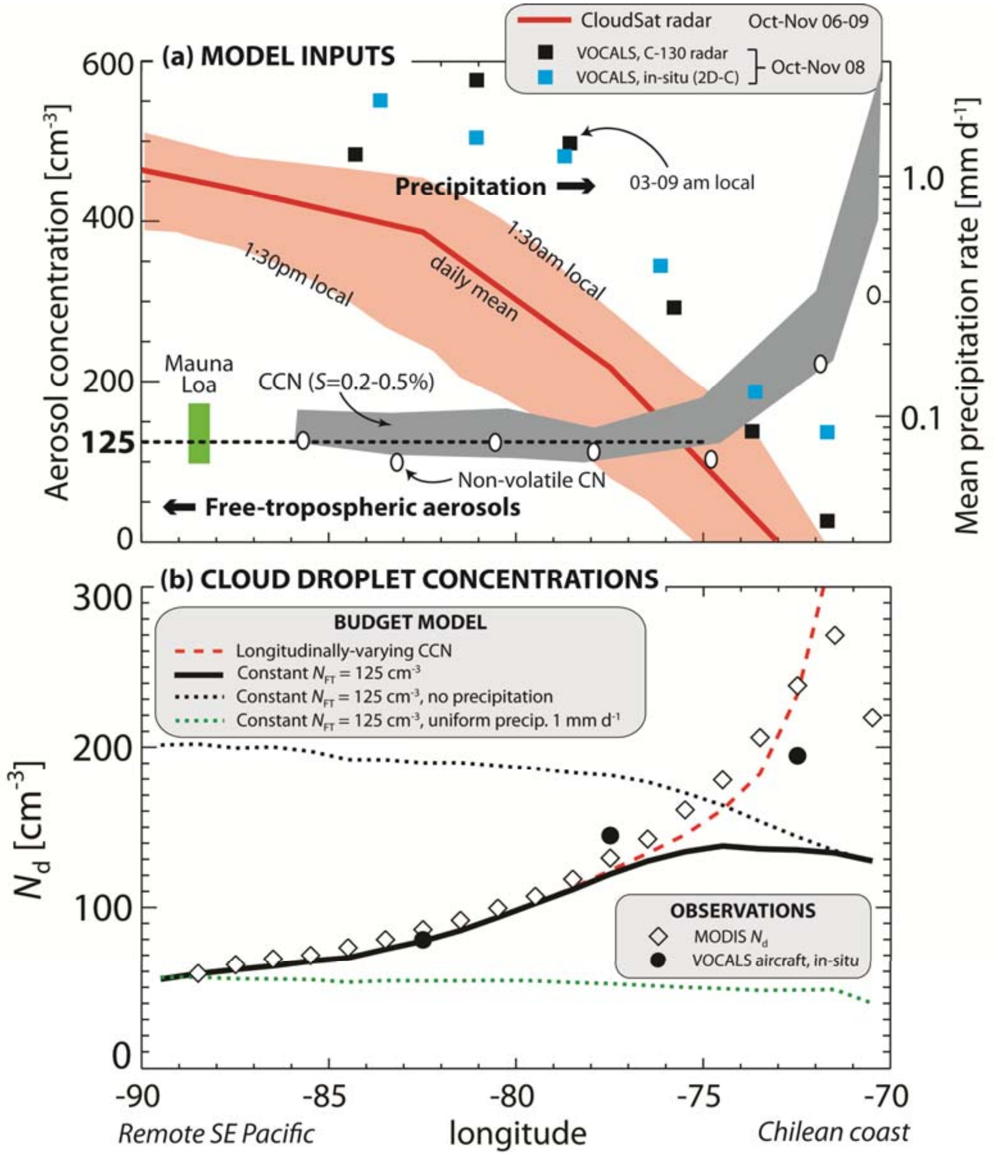
$$N_{eq} = \frac{\left(N_{FT} + \frac{F(\sigma)U_{10}^{3.41}}{Dz_i}\right)}{\left(1 + \frac{hKP_{CB}}{Dz_i}\right)}$$

Precip. Sink

- Concentration relaxes to FT concentration N_{FT} + wind speed dependent surface contribution dependent upon subsidence rate (Dz_i)
- Precipitation sink controlled by precipitation rate at cloud base P_{CB}. Use expression from Wood (2006).

Precipitation important in controlling gradient in cloud droplet concentration

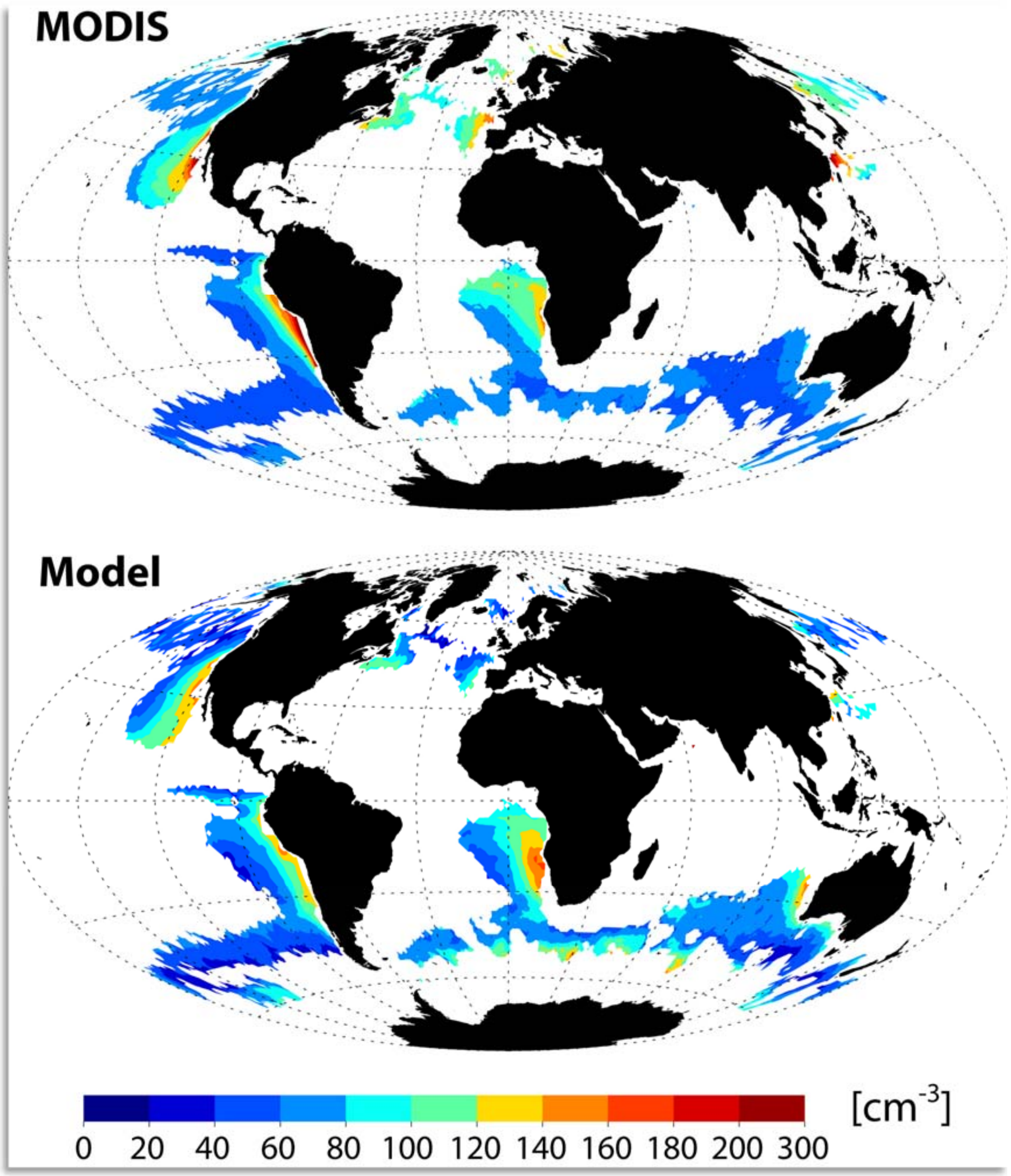
- Assume constant FT aerosol concentration
- Precipitation from CloudSat estimates from Lebsock and L'Ecuyer (2011)
- Observed surface winds
- Model N_d gradients mostly driven by precipitation sinks



Wood et al. (*J. Geophys. Res.* 2012)

Precipitation is primary control of N_d away from coastal zones

Model reproduces significant amount of variance in N_d over oceans $\Rightarrow$ implications for interpretation of AOD vs r_e relationships



Wood et al. (*J. Geophys. Res.* 2012)

Thoughts

- Much more work is required to interpret remotely-sensed AOD measurements as providing useful information about CCN
  - VOCALS region shows a doubling of Nd from the remote ocean to the coast but only a 10% increase in AOD
  - Differences explained by decreasing MBL depth and aerosol size
  - Need assimilation approaches, e.g. Saide et al. (2012)
- A large fraction of the variability in cloud droplet concentration over the remote oceans is driven by precipitation sinks as opposed to aerosol sources
  - Confounds interpretation of cloud vs aerosol relationships as indicative of aerosol indirect effects caused by anthropogenic pollution sources

Separating aerosol impacts from meteorological impacts on clouds

- Aerosol impacts on clouds are not simply explained by Twomey’s arguments

- Changes in macrophysical cloud properties produce radiative impacts of same order as those from Twomey (e.g. Lohmann and Feichter 2005, Isaksen et al. 2009)

Aerosol impacts on cloud

- An observed change in cloud property *C* is caused by changes due to meteorology *M* and aerosols *A*:

$$\delta C = \left(\frac{\partial C}{\partial M}\right)_A \delta M + \left(\frac{\partial C}{\partial A}\right)_M \delta A$$

meteorology-driven *aerosol-driven*

- To determine aerosol-driven changes on *C*, one needs to measure meteorology-driven changes
- This is a particularly arduous task, as the following examples demonstrate

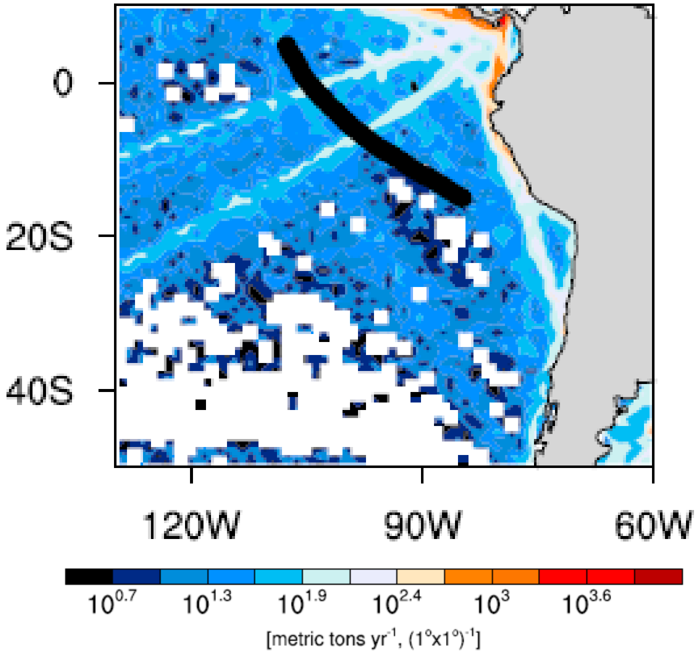
Stevens and Brenguier (2009)

Shiptracks

$$\delta C = \left(\frac{\partial C}{\partial M}\right)_A \underbrace{\cancel{\delta M}}_{= 0} + \left(\frac{\partial C}{\partial A}\right)_M \delta A$$

Shipping lanes

- Shipping emissions increase along preferred lanes
- **Control** clouds upstream; **perturbed** clouds downstream



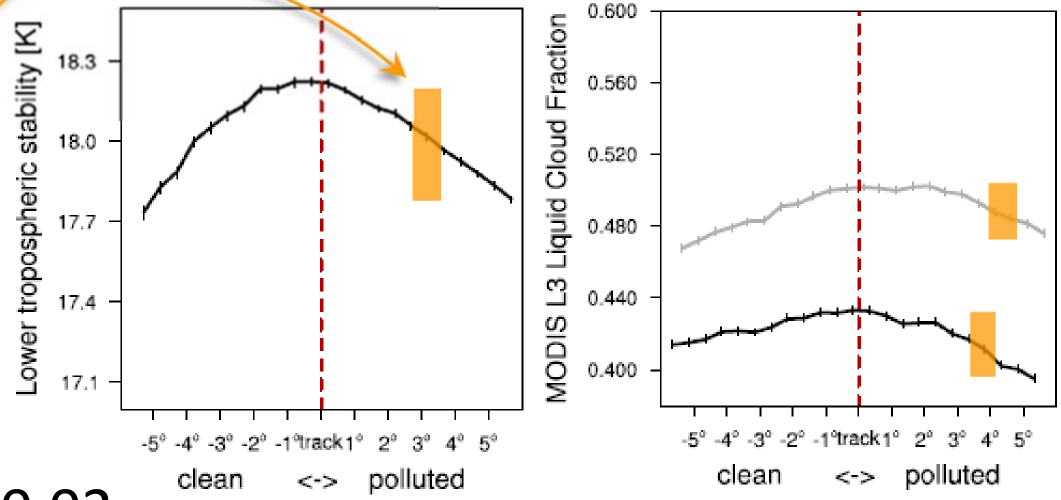
$$\delta f = \left(\frac{\partial f}{\partial LTS}\right)_A \delta LTS + \left(\frac{\partial f}{\partial A}\right)_M \delta A$$

Klein and Hartmann (1993)

$$\left(\frac{\partial f}{\partial LTS}\right)_A \delta LTS = 0.06\ \mathrm{K}^{-1} \times 0.4\ \mathrm{K}$$

$$= 0.024$$

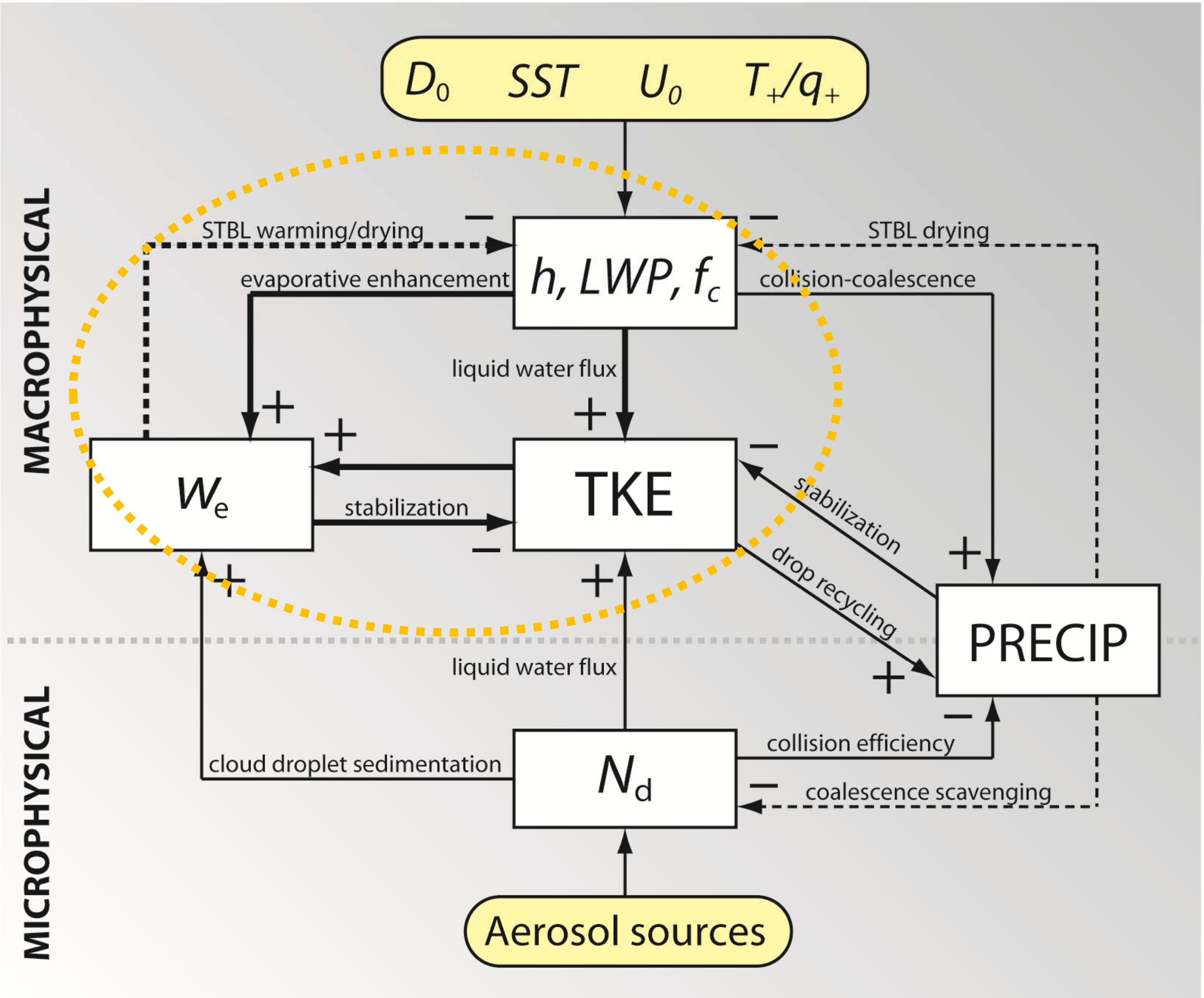
Observed $\delta f \approx 0.02\text{-}0.03$



A cloud cover increase of 0.02 represents a radiative forcing of 2 W m^{-2}

Peters et al. (ACP, 2011)

(Mostly) regulating feedbacks in stratocumulus


MACROPHYSICAL

MICROPHYSICAL

D_0 SST U_0 T_+/q_+

h, LWP, f_c

W_e

TKE

PRECIP

N_d

Aerosol sources

STBL warming/drying

STBL drying

evaporative enhancement

collision-coalescence

liquid water flux

stabilization

stabilization

drop recycling

liquid water flux

cloud droplet sedimentation

collision efficiency

coalescence scavenging